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LATE QUATERNARY GEOMORPHOLOGY OF THE GREAT SALT LAKE REGION,
UTAH, AND OTHER HYDROGRAPHICALLY CLOSED BASINS
IN THE WESTERN UNITED STATES: A SUMMARY OF OBSERVATIONS

By

Donald R. Currey

NASA Contract NAS5-28753, Final Report, Part III

Limneotectonics Laboratory Technical Report 89-3

July 25, 1989

Department of Geography

University of Utah

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Abstract

This report reviews attributes of Quaternary lakes and lake basins which are often important in the environmental prehistory of semideserts. Basin-floor and basin-closure morphometry have set limits on paleolake sizes; lake morphometry and basin drainage patterns have influenced lacustrine processes; and water and sediment loads have influenced basin neotectonics. Information regarding inundated, runoff-producing, and extra-basin spatial domains is acquired directly from the paleolake record, including the littoral morphostratigraphic record, and indirectly by reconstruction.

Increasingly detailed hypotheses regarding Lake Bonneville, the largest late Pleistocene paleolake in the Great Basin, are subjects for further testing and refinement. Oscillating transgression of Lake Bonneville began about 28,000 yr B.P., the highest stage occurred about 15,000 yr B.P., and termination occurred abruptly about 13,000 yr B.P. A final resurgence of perennial lakes probably occurred in many subbasins of the Great Basin between 11,000 and 10,000 yr B.P., when the highest stage of Great Salt Lake (successor to Lake Bonneville) developed the Gilbert shoreline. The highest post-Gilbert stage of Great Salt Lake, which has been one of the few permanent lakes in the Great Basin during Holocene time, probably occurred between 3,000 and 2,000 yr B.P.

A set of general observations regarding the nature of the paleolake record in semidesert basins is helping to guide Quaternary studies in the Great Basin.

Introduction

The geologic record of the Quaternary Period contains abundant evidence of past lakes (*paleolakes*) in semidesert basins (Morrison, 1968; Reeves, 1968; Street and Grove, 1979), including the many subbasins of the Great Basin, western U.S.A. (Snyder et al., 1964; Smith and Street-Perrott, 1983; Williams and Bedinger, 1984). The importance of Quaternary paleolakes in semidesert basins is two-fold: (1) paleolakes played significant, if transient, roles in shaping the floors and piedmonts of many basins (e.g., Mabbutt, 1977, pp. 180-214; Hardie et al., 1978; Eugster and Kelts, 1983), and (2) paleolakes encoded and stored, albeit episodically in many basins, enormous quantities of information concerning past environments (in Utah, e.g., Schreiber, 1958; Eardley and Gvosdetsky, 1960; Eardley et al., 1973; Spencer et al., 1984 and 1985; Baedeker, 1985). The purpose of this report is to review selected attributes of paleolakes and paleolake basins, as keys to retrieving and decoding paleolake-encoded and -stored paleoenvironmental information. This report draws particularly on lessons which are being learned from the paleolake record in the Great Basin, and especially from the records of Lake Bonneville and Great Salt Lake in the northeastern Great Basin (Figure 1).

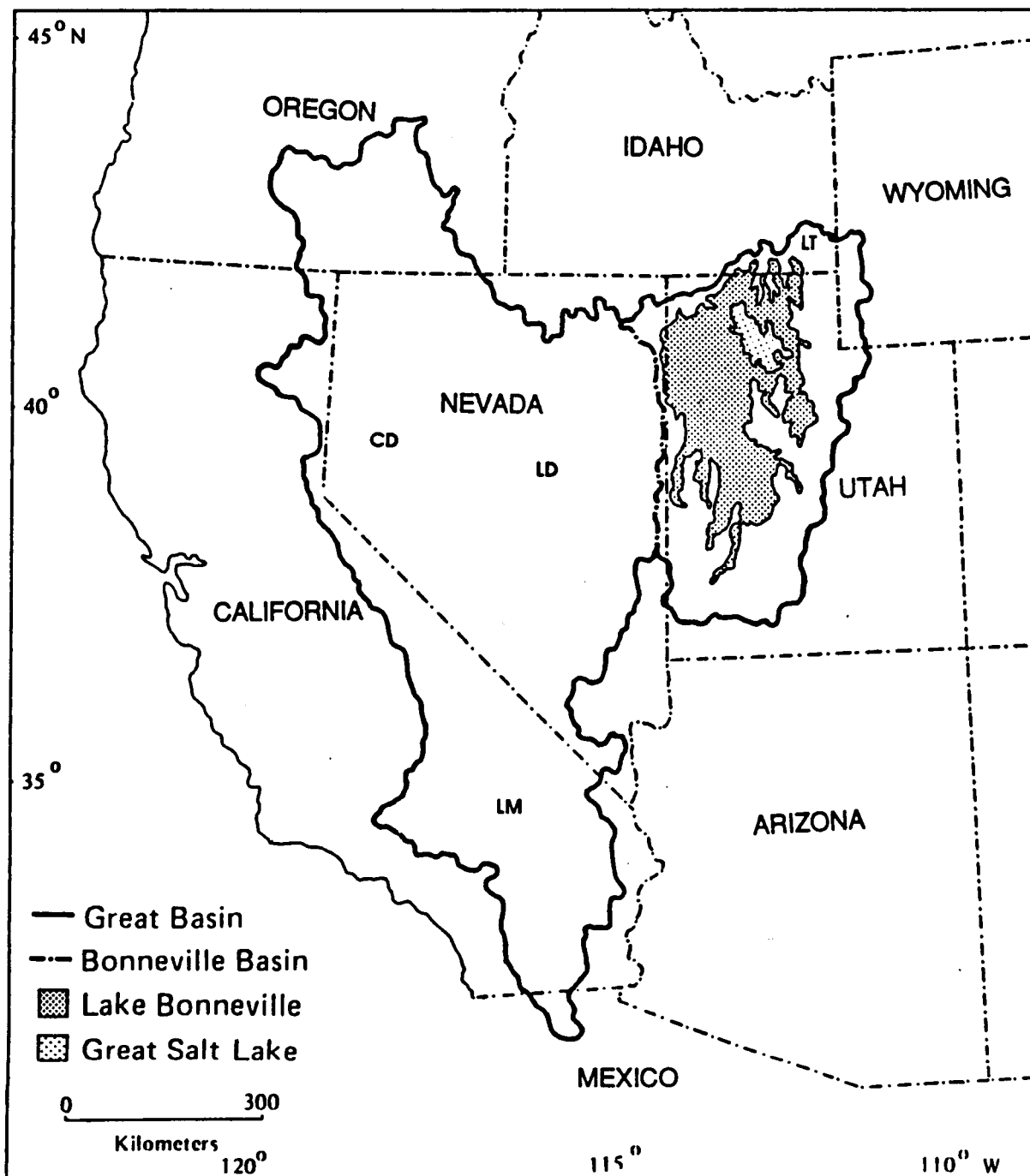


Figure 1. Locations of the Great Basin, Bonneville basin, Lake Bonneville, and Great Salt Lake in the southwestern U.S.A. (after Sack, 1989a). Other localities mentioned in the text are: *CD* = Carson Desert in the Lake Lahontan system of subbasins, *LD* = Lake Diamond, *LT* = Lake Thatcher, and *LM* = Lake Manix.

The episodic nature of paleolakes in semideserts has given rise to a long-standing tradition in which paleolake cycles have been represented diagrammatically as functions of relative or absolute time (e.g., Russell, 1885, figs. 31, 34, and 35; Gilbert, 1890, figs. 30 and 34; Morrison, 1965, figs. 2 and 4). In hydrographs, stratigraphic columns, correlation charts, and in histograms of paleolake occurrences through time, representations of paleolake cycles have appeared frequently in the Quaternary studies literature. Diagrams of paleolake cycles have tended to possess auras of authority, certitude, and spatiotemporal order which have beguiled workers in many Quaternary studies endeavors, including telecorrelation and global synthesis. However, as new data-gathering technologies have become available and as interpretive paradigms have been refined, the paleolake information which these diagrams convey has often been shown by subsequent workers (e.g., Scott, 1980; Oviatt, 1987 and 1988) to contain serious inaccuracies. This report outlines several perspectives by which paleolake reconstructions can be regarded as hypotheses which warrant systematic exploration, rather than scientific writ which warrants confident acceptance.

Paleohydrographic Continuum

Water bodies in semidesert basins have a wide range of possible *persistence* and *recurrence* through time (Figure 2). However, even where favorable geologic circumstances (Hutchinson, 1957, pp.

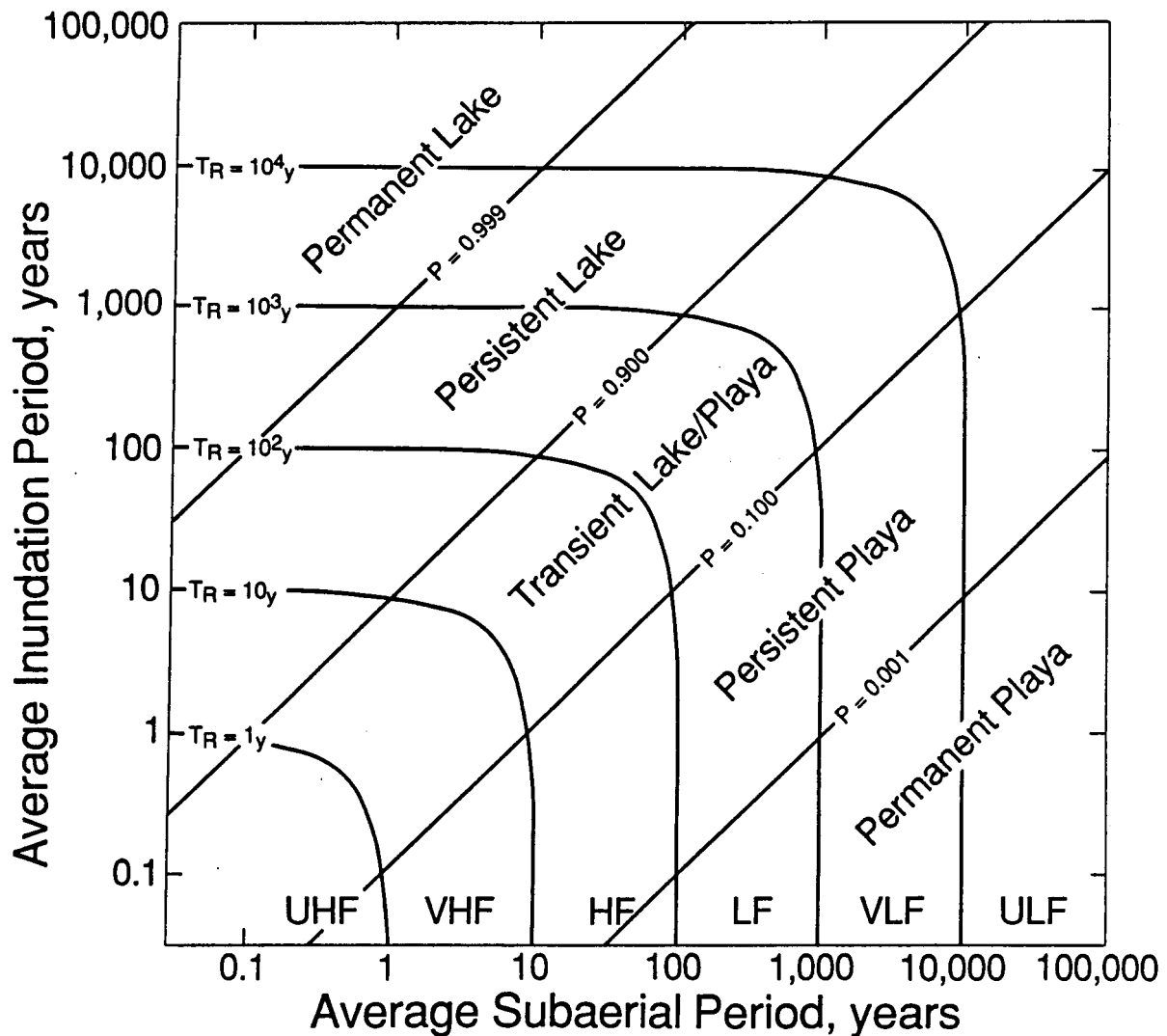


Figure 2. Persistence and recurrence of standing water on the floors of hydrographically closed basins. Persistence (P) is a dimensionless number from zero to one which expresses the average inundation period in years as a proportion of recurrence. Recurrence (T_R) is the time in years of the average inundation period plus the average subaerial period. Ultra high (UHF), very high (VHF), high (HF), low (LF), very low (VLF), and ultra low (ULF) frequency patterns of episodic inundation are characteristic of specific long-term intervals on the floors of some closed basins.

156-163) have provided suitable closed-basin hydrography (Figure 3), most semidesert basins have actually functioned as sensitive gauges of hydrologic mass balances (Street-Perrott and Harrison, 1985) and as efficient traps of lacustrine sediments (Hardie et al., 1978; Gwynn and Murphy, 1980; Eugster and Kelts, 1983; Smith et al., 1983) only within a limited range of hydrologic regimes (Langbein, 1961, pp. 2-4; Mifflin and Wheat, 1979, pp. 37-47)—and only within similarly limited intervals of geologic time.

For example, in many hydrographically closed semidesert basins the low stages of potential paleolakes have been well below the low-water limit of lacustrine processes, i.e., low-stage paleolake information has been off scale at the low end of available basin morphology (Figure 4, upper row). In some semidesert basins even the high stages of local water bodies have been off scale at the low end of the paleolake realm, i.e., have been in the paleopond-paleommarsh-paleoplaya realm (Figure 4, upper right). In other semidesert basins the highest paleolake stages have been off scale at the upper limit (*threshold*) of basin closure, with consequent loss of potential information because high-stage fluctuations were not completely recorded (Figure 4, left column). Temporally and spatially, the only complete records of paleolake history are encoded and stored in semidesert basins where the highest and lowest paleolake stages have been continuously on scale, below thresholds of basin closure and above the low-water limit of lacustrine processes (Figure 4, center).

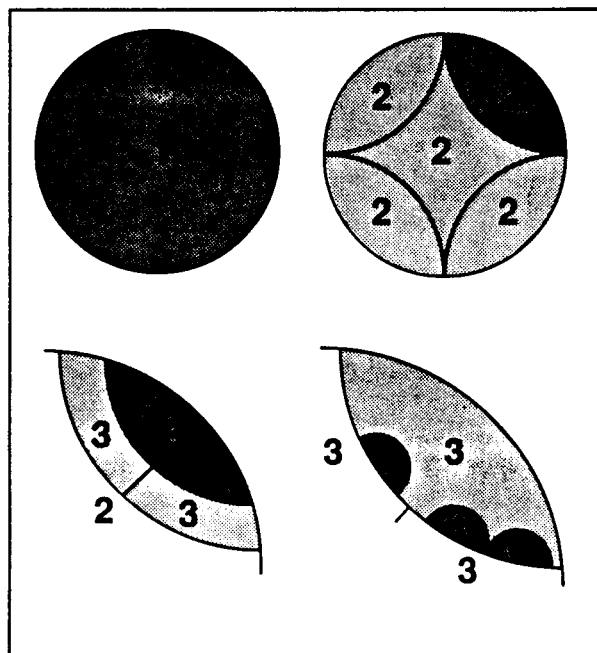


Figure 3. Schematic plan views of a nested hierarchy of hydrographically closed basins: 1 = first-order (e.g., Great Basin), 2 = second-order (e.g., Bonneville basin), 3 = third-order (e.g., Great Salt Lake, Great Salt Lake Desert, and Sevier Desert subbasins of Bonneville basin) and 4 = fourth-order (e.g., Puddle, Rush, and Cedar valley subbasins of Great Salt Lake subbasin).

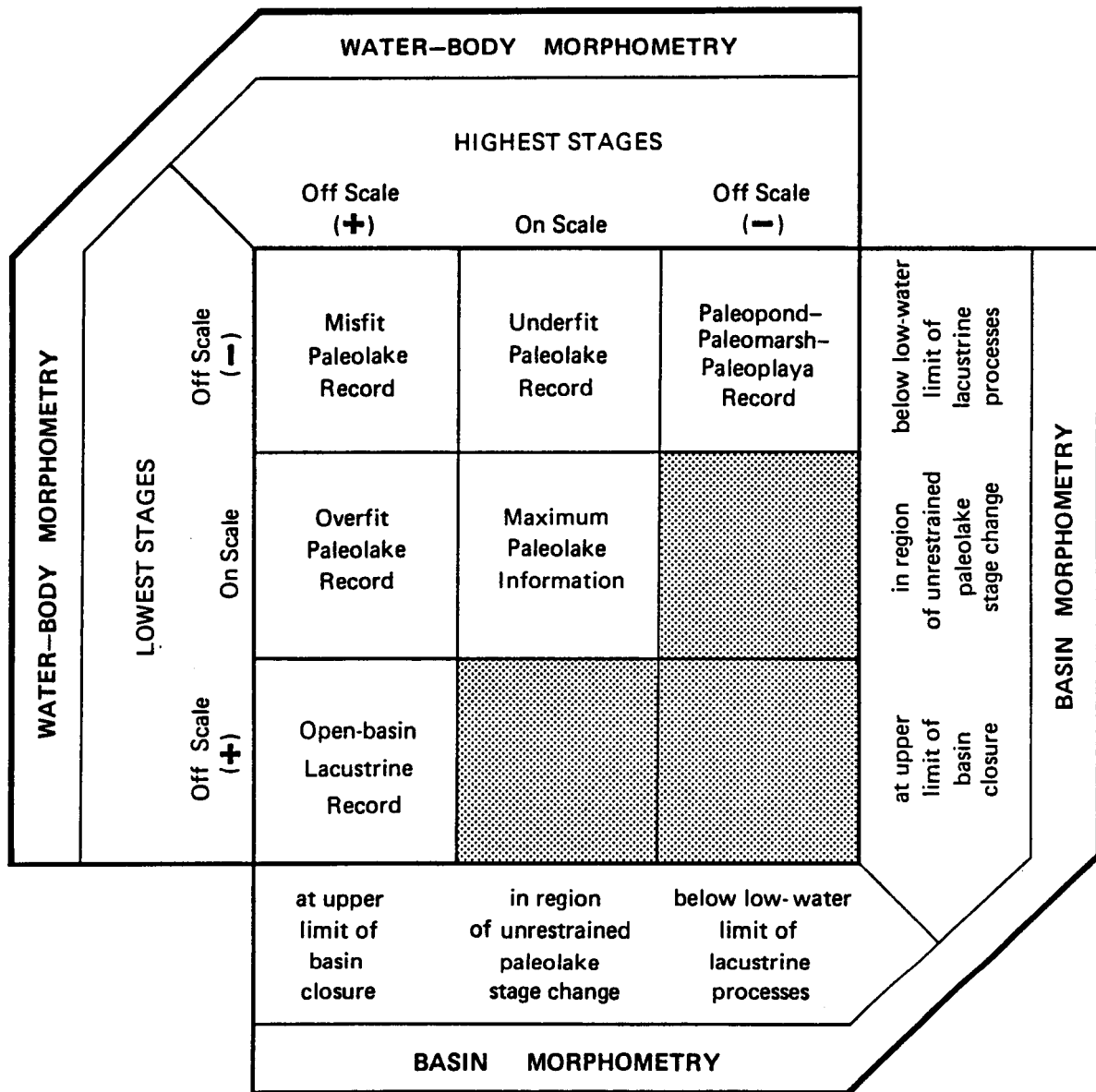


Figure 4. How well Quaternary geology recorded a particular interval of paleolake history is largely a function of how well the highest- and lowest-stage extremes of water-body morphometry fit available basin morphometry during that interval.

The hydrographic continuum which includes ponds, shallow lakes, and deeper lakes is difficult to subdivide by meaningful criteria (e.g., Burgis and Morris, 1987). As a generalization, lacustrine processes occur in, and are recorded by, water bodies deeper than 2 (Cowardin et al., 1979, pp. 11-12) to 4 m. In shallower water bodies: (1) aerodynamic turbulence is likely to cause frequent mixing of the water column and vigorous hydrodynamic stirring of offshore sediments, and (2) gently shelving topography is likely to suppress onshore propagation of waves in other than low energy bands of the wave spectrum, thus depriving the littoral zone of much of the sediment transport which is required to build beaches (Bascom, 1980).

The paleohydrographic continuum which is evident in the Quaternary record of Great Basin lowlands can be regarded as comprising four definable, but intergrading, paleohydrographic realms. *Paleolakes* were prehistoric perennial bodies of open standing water which were deep enough to leave well developed records of pelagial and littoral sedimentation. *Paleoponds* were prehistoric perennial or intermittent bodies of open standing water which were too shallow to leave well developed records of lacustrine sedimentation, but which were often effective agents of hydroaeolian planation—and of brim-full sedimentation in very small closed basins (e.g., Graf, 1989). *Paleomarshes* were prehistoric basin floors with hydromacrophyte wetlands, halomacrophyte meadows, saline mudflats, or saltflats (syn. = saltpans, evaporite crusts, salcretes, etc.)—and combinations

thereof, as in many modern saline "marshes" in Nevada (Papke, 1976)—which were wetted or dampened perennially by spring discharge or by evaporative pumping of shallow groundwater. *Paleoplayas* were prehistoric basin floors with relatively dry, salt-free, fine-grained surfaces—i.e., non-saline mudflats (claypans)—which were wetted briefly by rain or snowmelt, and which occasionally received significant runoff and accompanying sediments from surrounding piedmont and upland source areas. In essence, paleolakes and paleoponds were runoff-accumulating and -evaporating water bodies; paleomarshes were groundwater-discharging and -evaporating basin floors; and paleoplayas were stormwater-wetted and -evaporating basin floors.

Paleolake Containment

The morphometry of most semidesert basins is variable over long intervals of time. In addition to possible neotectonic deformation, geomorphic modification (Mabbutt, 1977) is virtually inevitable (Tables I and II). The capacity of a basin to have accommodated a range of potential paleolake sizes (Figure 5) depends largely on basin-closure and basin-floor morphometry. By lowering thresholds and raising floors, long-term changes in basin-closure and basin-floor morphometry have a tendency to diminish the range of potential paleolake sizes which a given basin can accommodate.

TABLE I

Basin-closure morphodynamics

Causes of basin-closure change	Threshold vertical change*
Geomorphic causes	
Piedmont alluvial fan accretion	+
Fluviodeltaic accretion	+
Loess accretion	+
Aeolian sand accretion	+
Aeolian deflation/abrasion	(-)
Overflow cryptoluvial piping/sapping	-
Overflow stream channel incising	-
Overflow stream channel head-cutting	-
Overflow flood channel scouring	-
Flood-induced kolk scouring	-
Flood-triggered landsliding	+
Overflow channel alluviation	+
Tectonic causes	
Far-field isostatic deflection	+ or -
Near-field hydro-isostatic deflection	+ or -
Near-field litho-isostatic deflection	(+) or -
Near-field glacio-isostatic deflection	(+) or (-)
Extensional seismotectonic displacement	+ or -
Transcurrent seismotectonic displacement	+ or -
Compressional seismotectonic displacement	(+) or (-)
Volcanic flow emplacement	+
Cinder cone construction	+
Tephra deposition	+

*Parentheses denote changes which may be of little significance in western North America. Symbols: + = threshold raised and - = threshold lowered.

TABLE II

Basin-floor morphodynamics

Causes of basin-floor change	Vertical change*	Horizontal change**
Geomorphic causes		
Proluvial (fan-toe sandflat) accretion	+	-
Deltoid (Mabbutt, 1977) deposition	+	0 or -
Deltaic deposition	+	0 or -
Pelagial deposition	+	0
Littoral erosion	0	0
Littoral deposition	+ or 0	0 or -
Evaporite deposition	+	+
Bog deposition	+	0
Hydroaeolian planation	approx. 0	+
Aeolian erosion	-	0 or -
Aeolian deposition	+	0 or -
Tectonic causes		
Far-field isostatic deflection	+ or -	0 or -
Near-field isostatic deflection	+ or -	0 or -
Seismotectonic displacement	+ or -	0 or -
Salt diapir doming	+	0 or -
Magma chamber inflation/deflation	+ or -	0 or -
Volcanic flow emplacement	+	-
Volcano construction	+	-
Volcano-tectonic subsidence	-	0 or -
Tephra deposition	+	0

*Symbols: + = basin floor raised, 0 = level of basin floor unchanged, and - = basin floor lowered.

**Symbols: + = basin floor enlarged, 0 = area of basin floor unchanged, and - = basin floor made smaller.

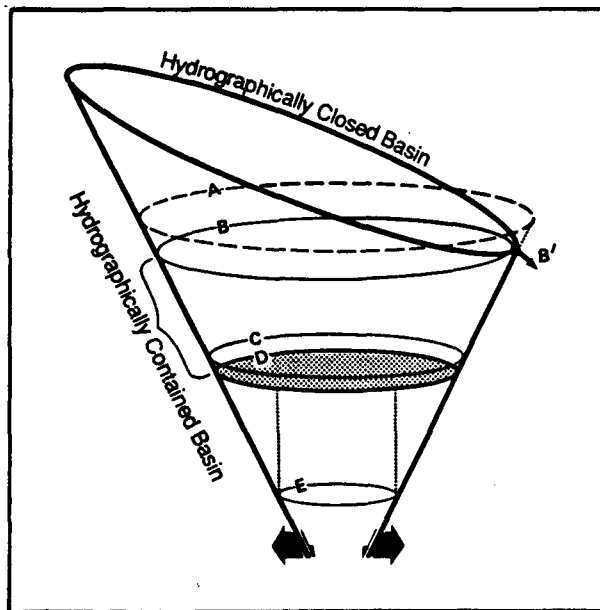


Figure 5. Dixie-cup, or "conical-graben," model of a paleolake basin, with a full spectrum of possible—and hydrologically potential but hydrographically impossible—paleolake stages, or "bathtub rings." A = hyperbasinal (*overfit*) stages inferred from hydrograph extrapolation; B = exobasinal (*overfit*) stage at upper limit of basin closure; B' = threshold of basin closure, subject to change from geomorphic and tectonic causes (Table I); B-C = endobasinal lacustrine stages; C = low-water limit of stages dominated by lacustrine processes; C-D = paleopond (*underfit*) stages dominated by hydroaeolian planation processes; D = paleomarsch or paleoplaya (*underfit*) stages, on basin floor subject to change from geomorphic and tectonic causes (Table II); and E = hypobasinal (*underfit*) stages inferred from hydrograph extrapolation.

In some semidesert basins—including those (Figure 1) of Lake Thatcher (Bright, 1963, pp. 115-121), Lake Bonneville (Gilbert, 1890, pp. 171-181; Malde, 1968; Currey, 1982, p. 23; Currey et al., 1984a; Jarrett and Malde, 1987; Smith et al., 1989, fig. 61), Lake Diamond (Hubbs et al., 1974, pp. 15-17; Mifflin and Wheat, 1979, fig. 7), and Lake Manix (Meek, 1989)—surface overflow under basin-full conditions caused thresholds of basin closure to be incised significantly (Table I), which has greatly reduced the capacity of those basins to contain lakes of relatively large size. Even in some basins without surface overflow, *cryptoluvial* sapping or piping associated with subsurface overflow—e.g., through the Stockton Bar at the north end of the Rush Valley subbasin of Lake Bonneville (Currey et al., 1983b, fig. 9)—caused thresholds of basin (and subbasin) closure to be degraded (Table I).

The floors of many semidesert basins have tended to aggrade and become larger in area through time (Table II), which has greatly reduced the capacity of those basins to contain lakes of relatively small size. In several subbasins of the Great Basin, including the Great Salt Lake Desert subbasin of the Bonneville basin and the Carson Desert (Figure 1) in the Lake Lahontan system of subbasins, basin floors have undergone progressive enlargement and overall flattening by *hydroaeolian planation*, a pattern of geomorphic development in which six sets of surficial processes occur in repetitive sequences on low-lying surfaces where shallow-water (depth < 4 m) and subaerial conditions alternate (Currey, 1987; Merola et al., 1989): (1) hydraulic erosion by wind-driven

bodies of shallow standing water entrains suspended sediment by subaqueous and, perhaps more importantly, by *circumaqueous* (lateral) scour, (2) hydraulic deposition of suspended sediment occurs as settling in still water and, perhaps more importantly, as *stranding* in evaporating water, (3) subaerial desiccation transforms newly deposited suspended sediment into clasts of curled and cracked dry mud, (4) aeolian erosion deflates mud clasts and abrades desiccated surfaces, (5) aeolian deposition of mud clasts occurs on *foredunes*, on upwind-opening *lunettes* which partially encircle deflated terrain, and on downwind-opening *antilunettes* which are encircled by deflated terrain, and (6) post-aeolian diagenesis sinters and cements aeolian mud clasts. Materials in (5) and (6) are recycled by repetitions of (1) *et seq.*

Playa expansion, playa-margin slope retreat, and playa smoothing by wind shifting of surface water have also been reported from the southwestern Great Basin (Motts, 1970). At localities throughout the Great Basin, the low-gradient sequences of *basin-flooring fine sediments* which are characteristic of recurrent hydroaeolian planation (Figure 6A) clearly differ in their *facies architecture* from the partially preserved, crudely concentric sequences of *basin-draping fine sediments* which are characteristic of recurrent paleolake cycles (Figure 6B).

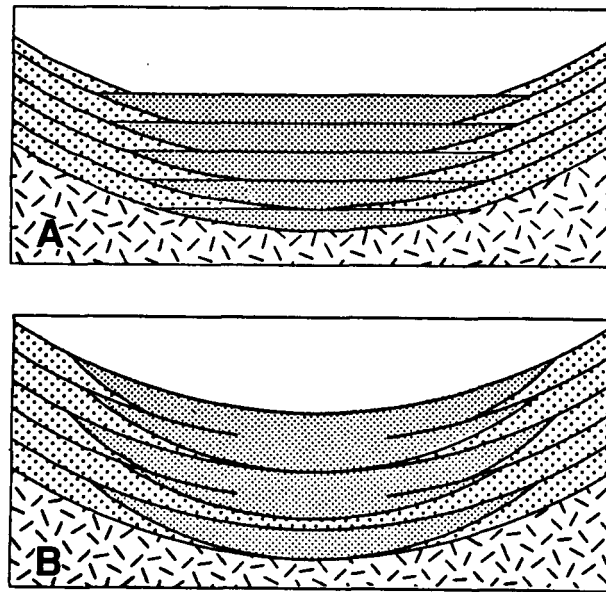


Figure 6. Idealized facies architecture resulting from repeated paleolake cycles which were (A) shallower than and (B) deeper than the variable critical depth which separates bottom erosion and transportation of silt-size materials from bottom accumulation of those materials (Håkanson and Jansson, 1983, pp. 177-204). Fine-grained bottom sediments and coarser piedmont sediments overlie bedrock.

Paleolake-Basin Interactions

Interactions between paleolakes and their basins have occurred in two main ways: (1) *basin hydrography* has tended to influence how paleolakes have functioned locally and basin-wide as lacustrine systems (Goldman and Horne, 1983; Wetzel, 1983), and (2) *water and sediment loads* imposed by large paleolakes have tended to influence basin neotectonics and, thereby, basin morphometry and shoreline geomorphology. The basic elements of basin hydrography are (1a) *lake morphometry* (Håkanson, 1981) and (1b) *basin drainage pattern*.

In dichotomous terms, lake morphometry pertains to the presence or absence of *strait-constricted arms* and/or *sill-* (submerged threshold) *constricted subbasins* in the inundated area of a closed basin (Table III). Basin drainage pattern pertains to the presence or absence of one or more major streams terminating in the inundated area of a closed basin (Table III). The influence of basin hydrography on paleolake processes tended to be least in *simple lakes* which lacked constrictions and major tributaries, and greatest in *compound-complex lakes* which had both constrictions and major tributaries (Table III). The compound-complex type includes the largest extant lake and the largest paleolakes in the Great Basin (Table III).

Some of the ways in which basin hydrography influences lacustrine processes can be outlined with reference to a hypothetical Great Basin paleolake (Figure 7). The *proximal (P)*, *medial (M)*, and *distal (D)* reaches of this paleolake have distinctive

TABLE III

Basic hydrographic elements of closed-basin lakes, with extant-lake (*) and palaeolake (**) examples from the Great Basin

Basin drainage pattern	Lake morphometry	
	Lakes without constricted arm(s) or closed subbasin(s)	Lakes with constricted arm(s) and/or closed subbasin(s)
Lakes without inflow from major river(s)	simple lakes Summer Lake* palaeolakes common	compound lakes extant lakes rare Lake Chewaucan**
Lakes with inflow from major river(s)	complex lakes Sevier Lake* Owens Lake* Lake Thatcher**	compound-complex lakes Great Salt Lake* Lake Searles** Lake Bonneville** Lake Lahontan**

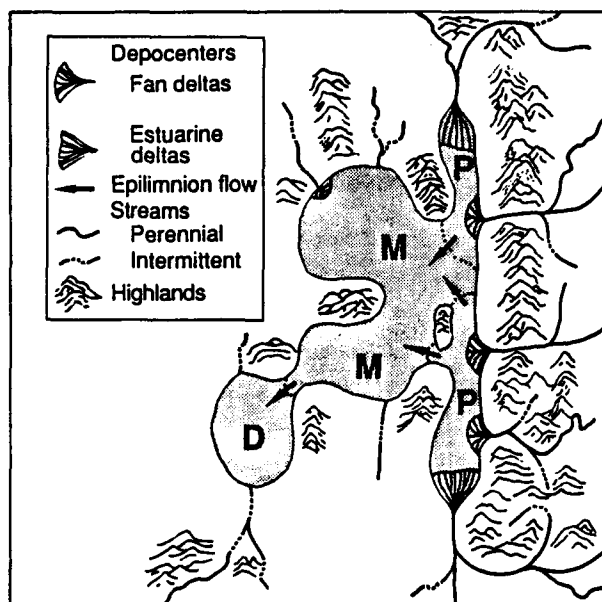


Figure 7. Hypothetical paleolake, showing general locations of proximal (P), medial (M), and distal (D) reaches. Arrows indicate directions of epilimnion net flow. Paleolake dynamics which typify the proximal, medial, and distal reaches are outlined in Figure 8.

sets of hydrologic, hydrodynamic, depositional, and lithofacies characteristics (Table IV). In semiquantitative terms, the proximal, medial, and distal reaches of this paleolake can be regarded as distinctive loci in ternary fields which depict inputs to the lacustrine water balance (Figure 8A), origins of horizontal motion in the epilimnion (Figure 8B), and origins of sediment at the bottom of the water column (Figure 8C).

Episodic waxing and waning of Pleistocene water loads caused significant *hydro-isostatic deflection* in the basins of large paleolakes, including Lake Lahontan (Mifflin and Wheat, 1971) and Lake Bonneville (Gilbert, 1890, pp. 362-392; Crittenden, 1963; Currey, 1982; Bills and May, 1987). Hydro-isostatic deflection originating in the Bonneville basin even distorted shoreline development in other subbasins of the eastern Great Basin (Currey et al., 1984b; Bills et al., 1987). In some paleolake basins sediment loads have probably caused *litho-isostatic deflection*, particularly at and near deltaic depocenters (Currey, 1982, p. 25 and pl. 1).

Where semidesert basins are largely of tectonic origin—as in the Basin and Range extensional tectonic province, which includes most of the Great Basin (Thornbury, 1965, pp. 471-505; Hunt, 1974, pp. 480-535; Eaton, 1982; Fiero, 1986)—total neotectonic deformation is the sum of seismotectonic displacement and isostatic deflection (Figure 9). Ongoing research in the Bonneville basin is uncovering mounting evidence that the vertical direction (whether subsidence or rebound) and tempo of isostatic deflection, by affecting regional patterns of stress in the lithosphere, played

TABLE IV

Palaeolimnology of proximal, medial, and distal palaeolake reaches

Proximal reach	
Hydrology	local runoff was dominant source of water balance input and epilimnion salinity was relatively low
Hydrodynamics	water-balance-driven outflow prevailed in epilimnion
Depodynamics	terrigenous sedimentation prevailed
Lithofacies	fluviodeltaic clastics are prevalent
Medial reach	
Hydrology	transbasin flow was important source of water balance input
Hydrodynamics	water-balance-driven transbasin flow and wind-driven flow prevailed in epilimnion
Depodynamics	limnogenous sedimentation tended to prevail
Lithofacies	pelagial micrite (typically calcite) and littoral coarse clastics are prevalent
Distal reach	
Hydrology	transbasin flow was dominant source of water balance input and epilimnion salinity was relatively high
Hydrodynamics	water-balance-driven inflow prevailed in epilimnion
Depodynamics	limnogenous sedimentation prevailed
Lithofacies	pelagial micrite (typically aragonite) and littoral carbonates are prevalent

Figure 8. Loci of proximal (P), medial (M), and distal (D) paleolake reaches in ternary fields depicting paleolake dynamics.

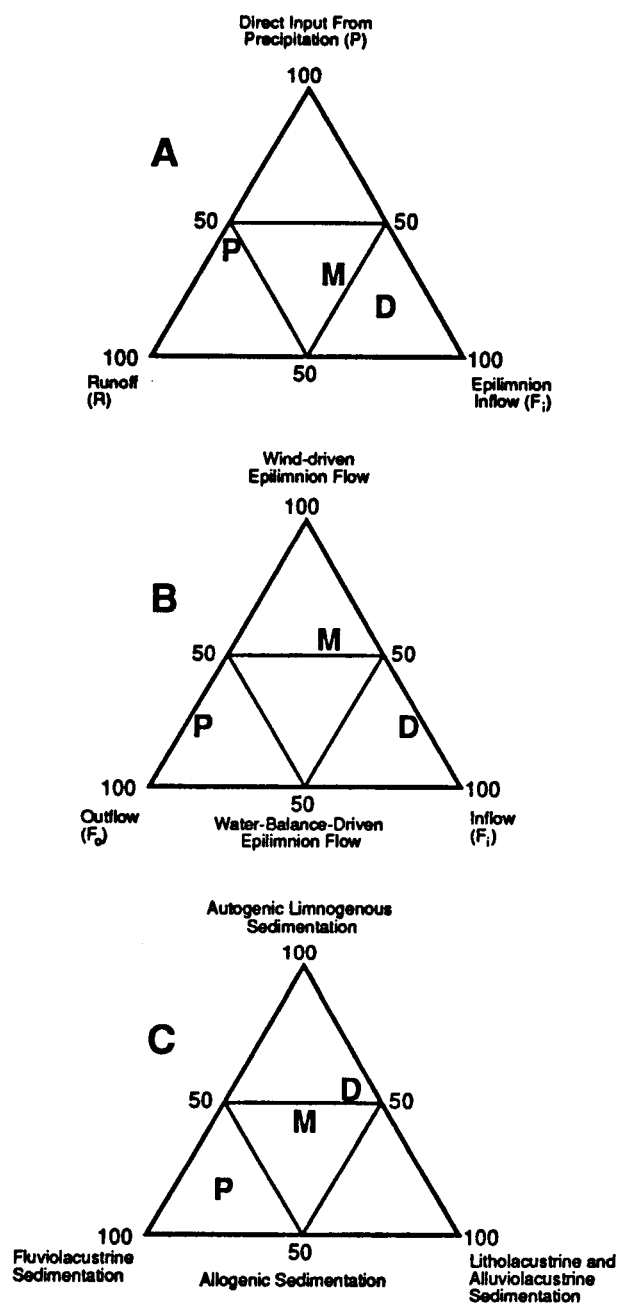
(A) Water balance inputs, as percentages of long-term total input to a paleolake reach; water balance equation of continuity in a paleolake reach is $(P + R + F_i) - (E + F_o) = \Delta S$, where P is direct precipitation, R is runoff from adjacent drainage area, F_i is inflowing epilimnion water, E is evaporation, F_o is outflowing epilimnion water, and ΔS is change in storage (change in lake stage).

(B) Sources of horizontal motion in epilimnion, as percentages of long-term total epilimnion flow within a paleolake reach; net flow across a paleolake reach is $F_i - F_o$.

(C) Sources of sediment at bottom of water column, as percentages of long-term total sedimentation. *Limnogenous sediments* comprise all materials which originate physiochemically or biochemically within or at the bottom of the water column, irrespective of depth. *Fluviolacustrine sediments* comprise all clastic materials, including glacial outwash, which are introduced by streams—and which in part are widely dispersed offshore and in part are localized nearshore, often in low-gradient suspended-load deltas, bed-load fan deltas, poorly sorted underflow fans, and aggraded-prograded estuaries. *Litholacustrine sediments* comprise all clastic materials which are derived from erosion of cliffs and shore platforms in coastal bedrock (Trenhaile, 1987). *Alluvio-lacustrine sediments* comprise all clastic materials which are derived from erosion of bluffs in piedmont alluvium.

Figure 8

(For explanation, see caption on previous page.)



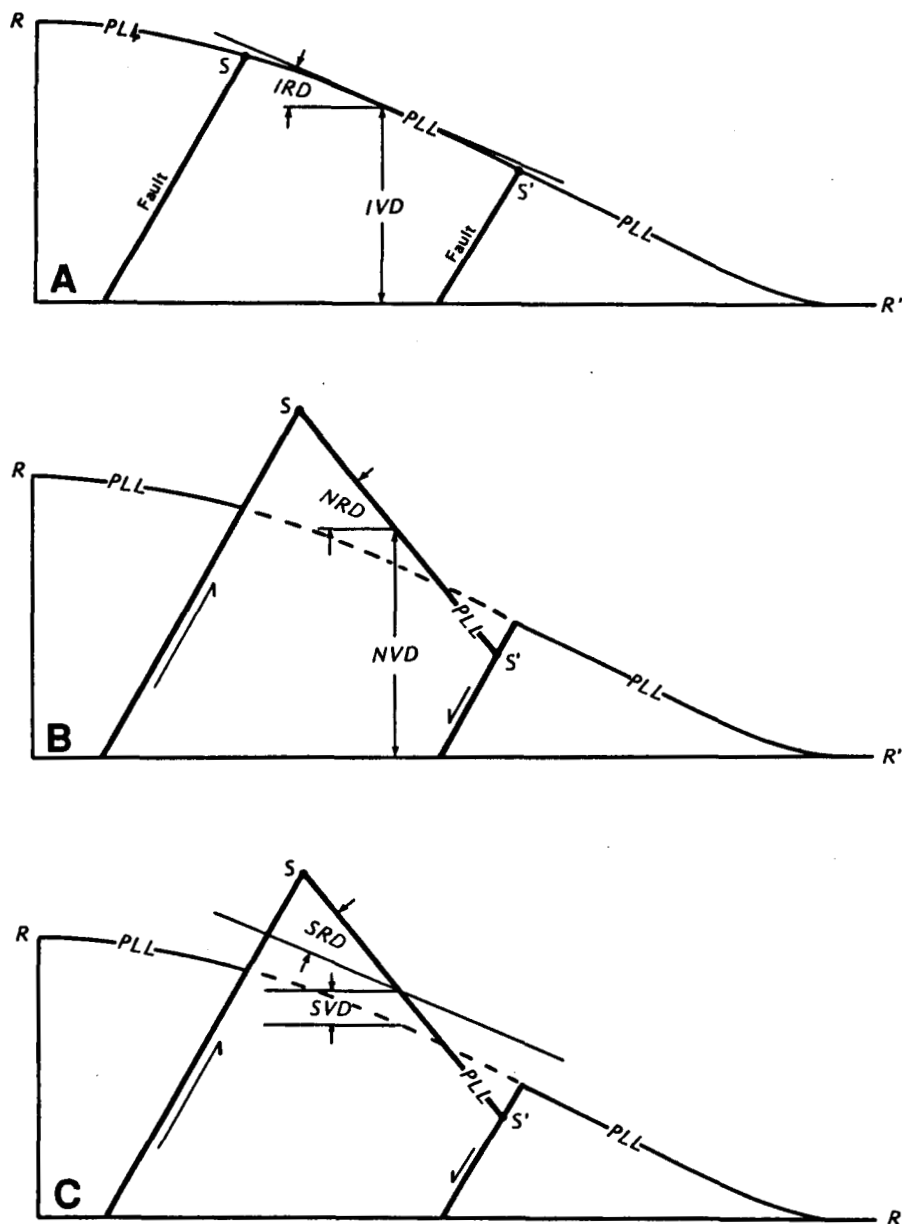


Figure 9. Schematic profiles showing regional and subregional deformation of an originally horizontal paleolake level (PLL) of known age (Currey, 1988b, fig. 2). $R-R'$ = regional trend of an isostatically deflected PLL. $S-S'$ = fault-bounded subregional PLL segment in which are depicted: (A) isostatic vertical deflection (IVD) and isostatic rotational deflection (IRD) components of total isostatic deflection; (B) neotectonic vertical deformation (NVD) and neotectonic rotational deformation (NRD) components of total neotectonic deformation; and (C) seismotectonic vertical displacement (SVD) and seismotectonic rotational displacement (SRD) components of total seismotectonic displacement. Neotectonic deformation is the sum of isostatic deflection and seismotectonic displacement.

a significant role in modulating the tempo of seismotectonic events. Neotectonic complexity clearly contributes to the complexity of the paleolake record in some semidesert basins; conversely, detailed studies of paleolake records offer excellent opportunities to decipher neotectonic history (Currey, 1988b).

Paleolake Record

The assortment of Quaternary lakes, lacustrine cycles, shoreline chronologies, and paleoenvironmental interpretations which have been hypothesized from fragments of the paleolake record in the western United States is large and varied (Feth, 1964; Morrison, 1965; Mehringer, 1977 and 1986; Smith and Street-Perrott, 1983; Benson and Thompson, 1987a; Benson et al., in press). Some hypotheses regarding paleolakes have grown in credibility with each iteration of testing and refinement, some have been falsified conclusively, and some have become objects of protracted debate (e.g., Van Horn, 1988; Van Horn and Varnes, 1988), sometimes by eluding falsification through lack of specificity. Experience has shown that the viability of a specific paleolake hypothesis tends to be directly proportional to the scope of the paleolake record on which it is based. That is, the most viable hypotheses, or most probable paleolake predictive syntheses, tend to be those which are distilled from multiple channels of basin-wide information.

Paleolake information belongs to three spatial domains: (1) the physically *inundated area* of a paleolake basin, (2) the runoff-

producing *drainage area* of a basin, and (3) the *extra-basin region* which can influence a basin atmospherically and geologically (Figure 10). Primary paleolake information is acquired directly from the physical record (Figure 10, bold boxes). Reconstructed (secondary) paleolake information—such as salinity, sediment budget, circulation, oxygenation, stratification, trophic status, temperature, water budget, and local and regional climate—is acquired by deductive reasoning, often with the aid of numerical models, which starts from a foundation of primary information and proceeds counter to the directions of causality which are represented by arrows in Figure 10. Experience in the Great Basin suggests that the most satisfying paleolake reconstructions proceed from primary information which spans the full range of proximal, medial, and distal reaches, and the full range of pelagial and littoral paleoenvironments. Experience also has shown that analyzing the *morphostratigraphy* of the littoral record, through the combined methods of geomorphology and stratigraphy (Figure 11), is an effective strategy for retrieving some of the more useful information in the paleolake record (see Concluding Observations 5 through 11).

Lake Bonneville Hypotheses

An extensive paleolake record in the Bonneville basin provides a wealth of primary information which is being used by a growing cohort of workers to build and test an evolving system of inter-

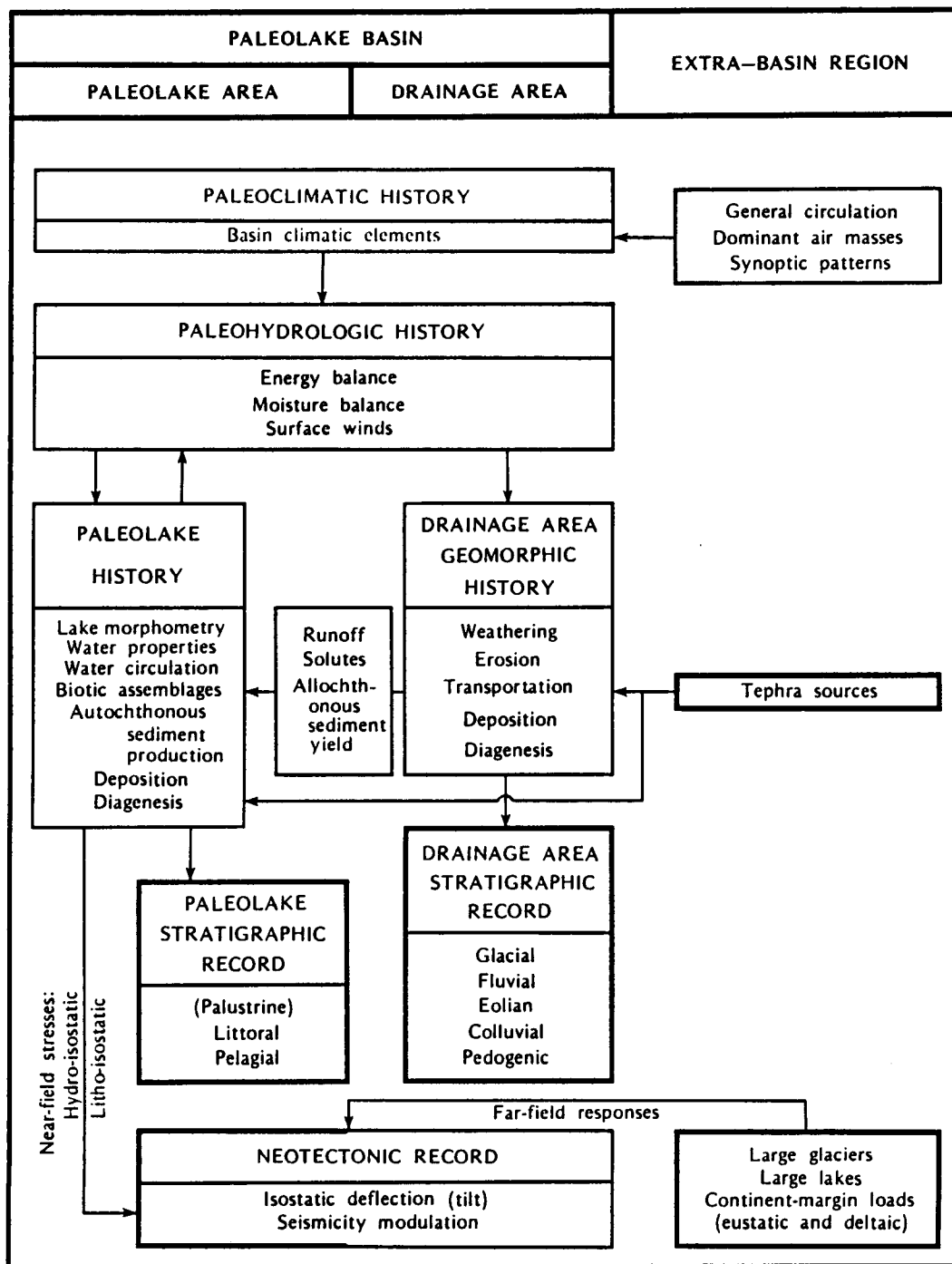


Figure 10. Paleolake history is reconstructed from the paleolake and regional geologic records (bold boxes) by *response-process reasoning* which is essentially counter to process-response causality (arrows).

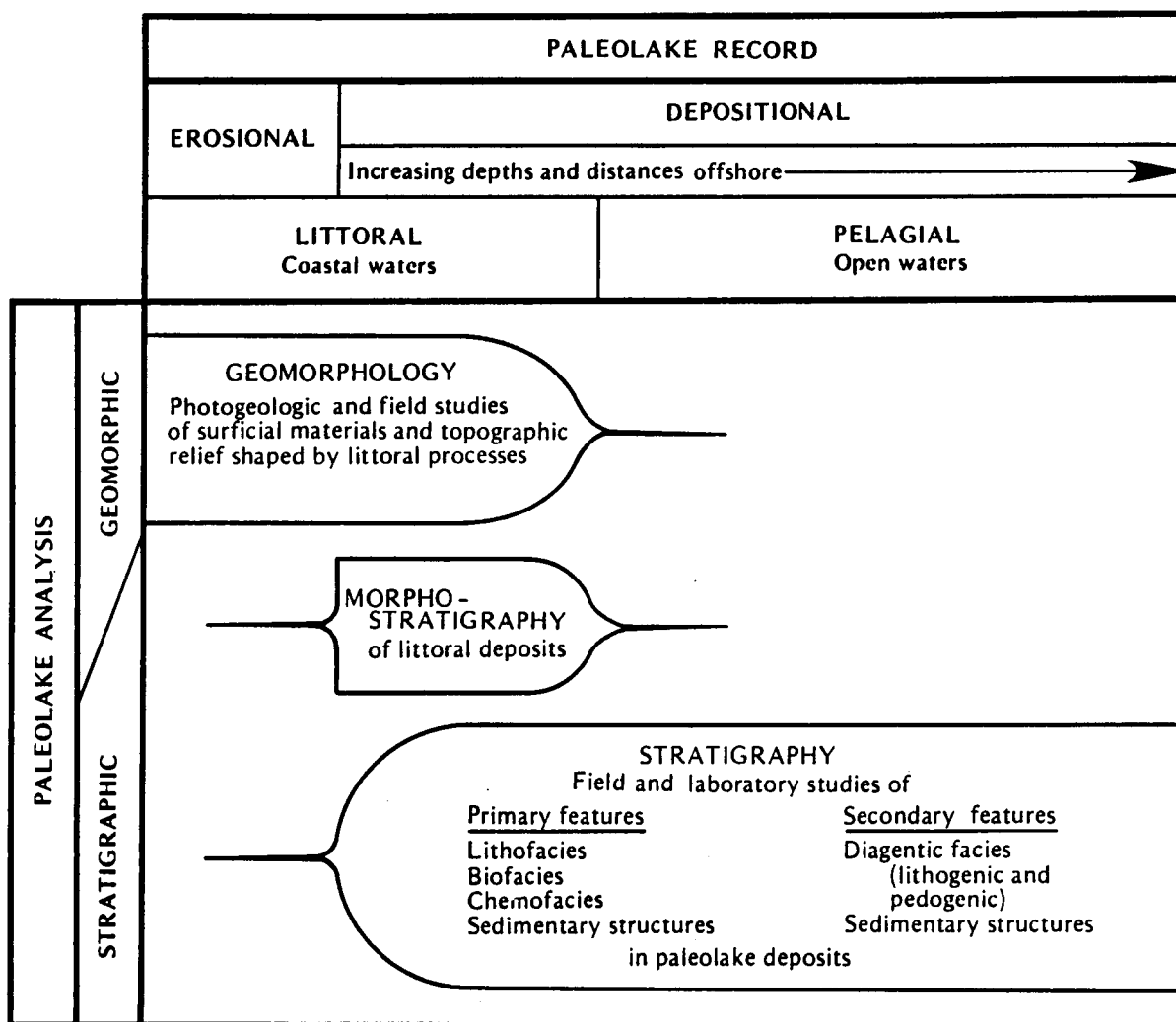


Figure 11. Morphostratigraphic analysis of the littoral depositional record uses the methods of geomorphology and stratigraphy to extract vital paleolake information (Currey and Burr, 1988, fig. 1).

related paleolake hypotheses (e.g., Currey et al., 1983a; Scott et al., 1983; Spencer, 1983; Currey et al., 1984a and 1984b; Currey and Oviatt, 1985; McCoy, 1987; Oviatt, 1987; Oviatt and Currey, 1987; Oviatt et al., 1987; Currey and Burr, 1988; Machette and Scott, 1988; Oviatt, 1988; Sack, 1989a). Several predictive syntheses which are based mainly on morphostratigraphic analysis of the littoral record in the Lake Bonneville-Great Salt Lake region are regarded here as subjects for further testing and refinement.

The last deep-lake cycle (Bonneville paleolake cycle) in the Bonneville basin occurred during the interval from about 28,000 to 13,000 yr B.P., which was essentially synchronous with oxygen isotope stage 2 (Shackleton and Opdyke, 1973) and chronostratigraphic interval 3.1-2.0 (Martinson et al., 1987) of the marine record. As generalized in Figure 12, the cycle comprised three major phases: (1) a protracted phase of closed-basin, oscillatory-transgressive stages—interrupted by an important regression between 21,000 and 20,000 yr B.P. (Stansbury oscillation of Oviatt et al., in press)—until about 15,300 yr B.P.; (2) a phase of intermittently open-basin, threshold-controlled stages—interrupted by an important regression between 15,000 and 14,500 yr B.P. (Keg Mountain oscillation of Currey et al., 1983b) and highlighted by catastrophic downcutting of the threshold about 14,500 yr B.P. (Bonneville Flood of Malde, 1968; Jarrett and Malde, 1987)—from about 15,300 to 14,200 yr B.P.; and (3) a brief phase of closed-basin, rapidly regressing stages after about 14,200 yr

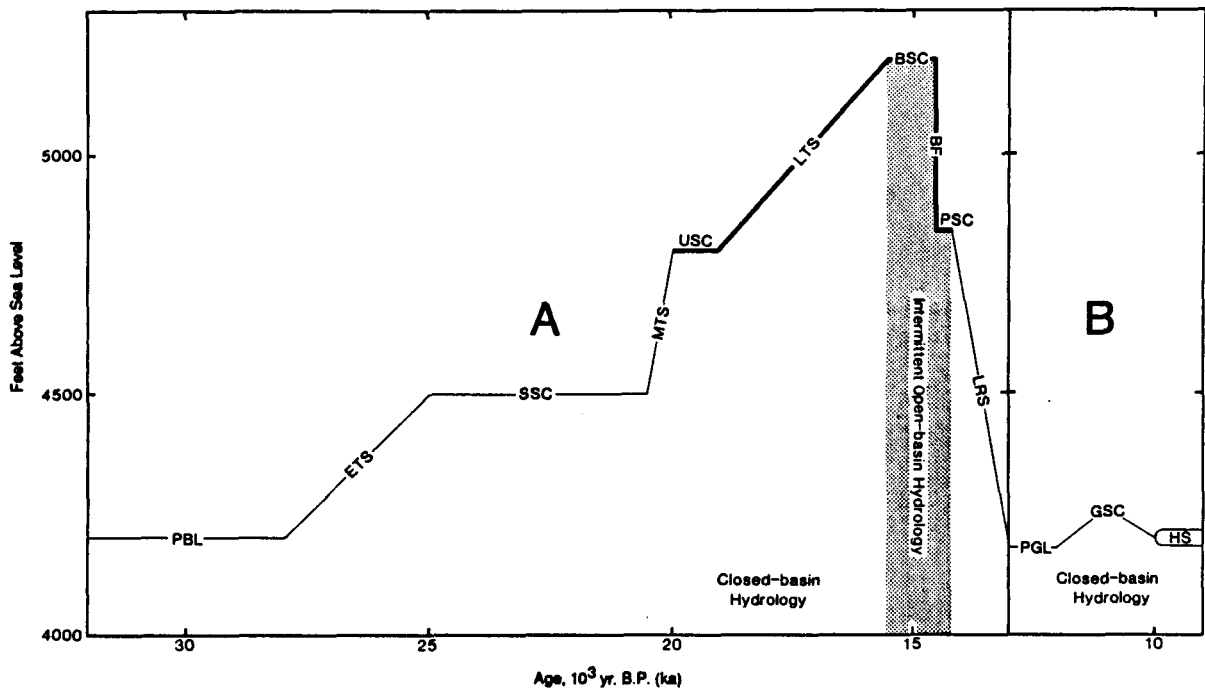


Figure 12. Schematic hydrograph of the Bonneville basin during the Bonneville paleolake cycle (A) and in early post-Bonneville time (B). Hydrograph segments are: *PBL* = pre-Bonneville low, *ETS* = early transgressive stages, *SSC* = Stansbury shoreline complex, *MTS* = middle transgressive stages, *USC* = unnamed shoreline complex, *LTS* = late transgressive stages, *BSC* = Bonneville shoreline complex, *BF* = Bonneville Flood, *PSC* = Provo shoreline complex, *LRS* = late regressive stages, *PGL* = pre-Gilbert low, *GSC* = Gilbert shoreline complex, and *HS* = Holocene stages. Heavy line denotes spatiotemporal range of the littoral record at the Stockton Bar, a classic locality 25 km south of Great Salt Lake (Burr and Currey, 1988, fig. 3).

B.P.

Paleogeography of the pre-Flood (Bonneville) and post-Flood (Provo) open-basin stages of Lake Bonneville, as well as of the subsequent (Gilbert) highest stage of Great Salt Lake, is outlined in Figure 13. In Figure 14, profiles through the center of the Bonneville basin from south-southwest to north-northeast depict the cumulative hydro-isostatic deflection which has occurred since the paleolake stages shown in Figures 12 and 13.

A detailed reconstruction of the intermittently open-basin phase (Figure 12, shaded column) has been proposed in a linear model of threshold-controlled shorelines of Lake Bonneville (Currey and Burr, 1988). Age estimates in the linear model (Figure 15) reflect refinements of the chronology of Currey and Oviatt (1985), and have an average error which probably does not exceed 300 ^{14}C years. Details of the linear model (which assumes that at any locality the rate of hydro-isostatic subsidence was constant before and after the Keg Mountain oscillation prior to the Flood, and the rate of hydro-isostatic rebound was constant subsequent to the Flood) provide a means of exploring the ways in which changing lake stages (hydrographic kinematics) and the deflecting basin (isostatic kinematics) interacted with each other and with changing threshold geomorphology (geomorphic kinematics) to develop the littoral morphostratigraphic signatures which are observed from locality to locality in the Bonneville basin. The model is testable by at least three lines of evidence, viz., hypsometric, chronometric, and morphostratigraphic. With refinements in

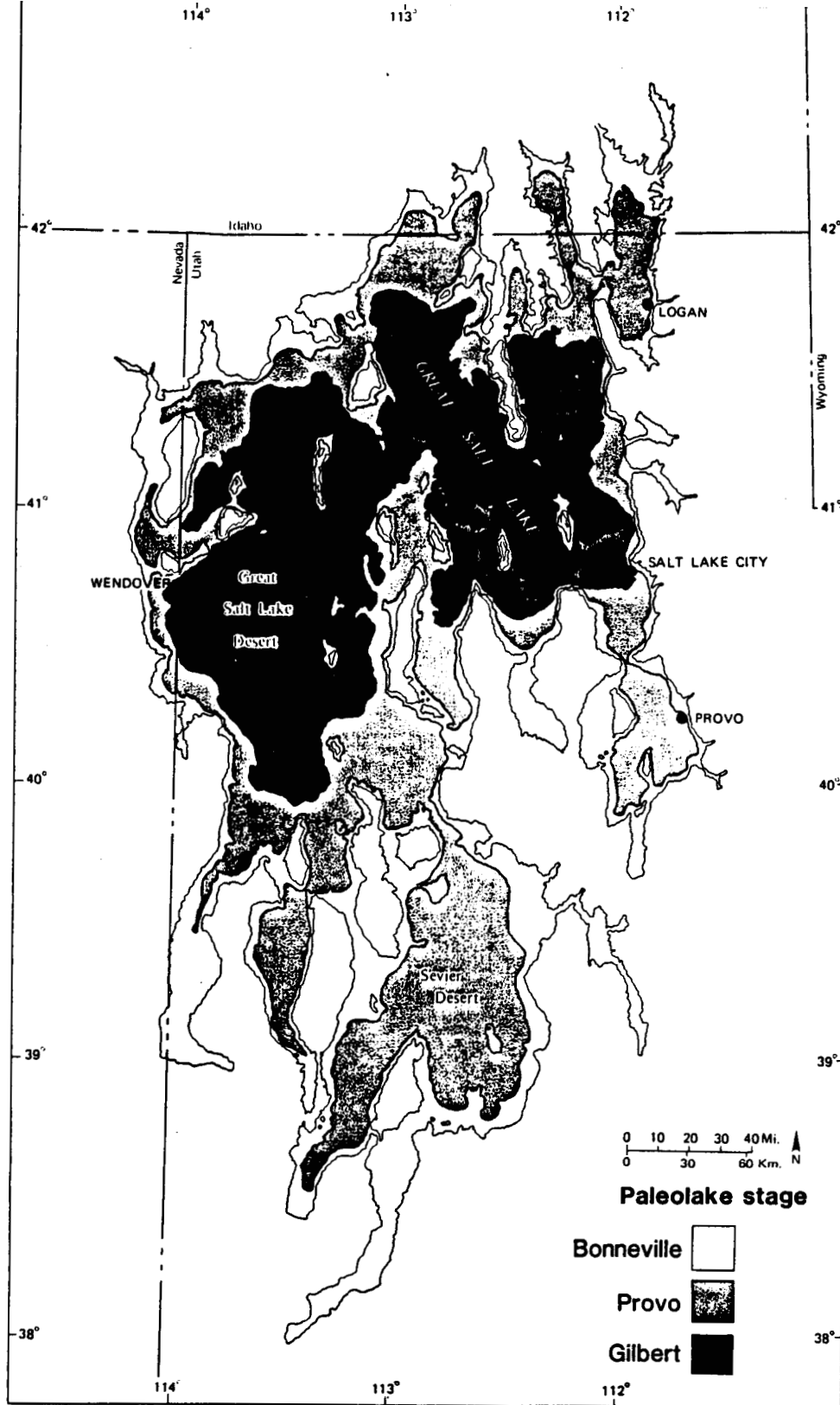


Figure 13. Map of the Lake Bonneville region (adapted from Currey et al., 1984a, figs. 1 and 2) depicting the open-basin (Bonneville) stage prior to the Bonneville Flood, the open-basin (Provo) stage subsequent to the Flood, and the highest (Gilbert) stage of Great Salt Lake subsequent to the final regression of Lake Bonneville.

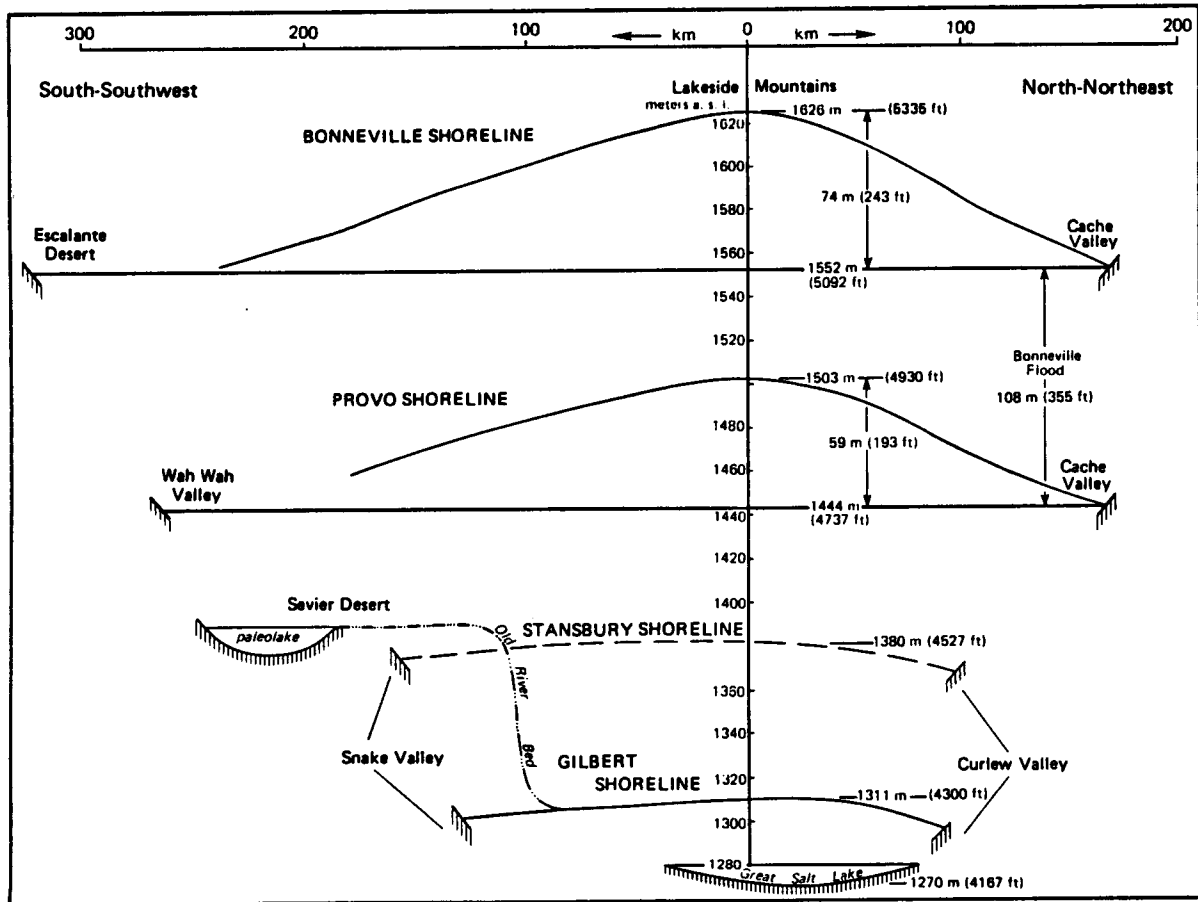


Figure 14. Regional hydro-isostatic deflection of major Bonneville basin shorelines depicted in Figures 12 and 13 (adapted from Currey, 1982; Currey, 1988b, fig. 1).

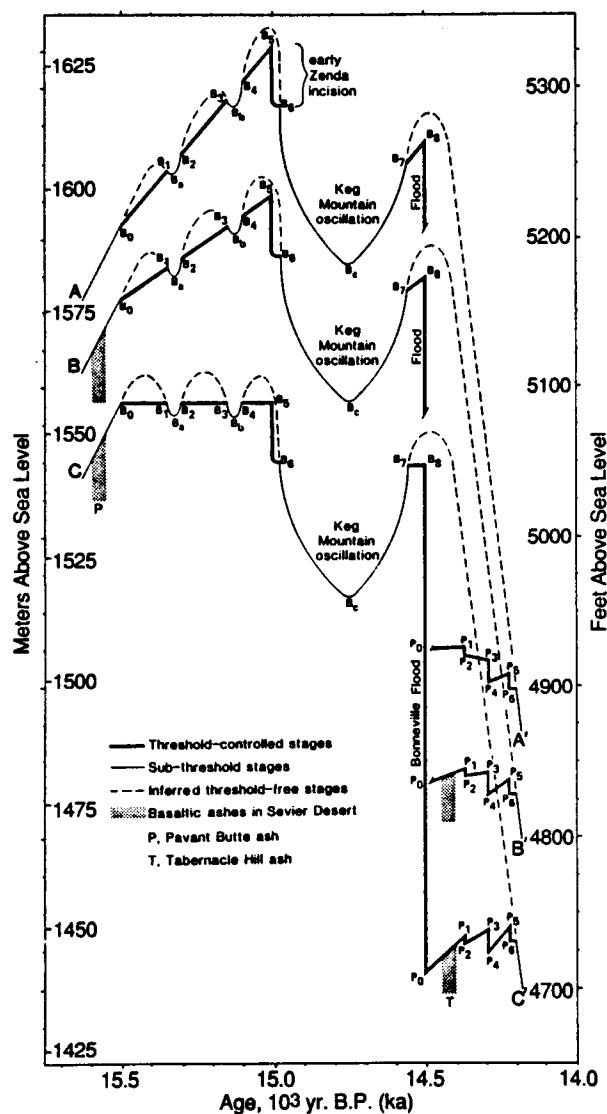


Figure 15. Linear model of modern altitudes of threshold-controlled stages of the Bonneville paleolake cycle: A-A' = at or very near the basin centroid of greatest hydro-isostatic deflection; B-B' = at Stockton Bar, in a basin-interior area of intermediate deflection; and C-C' = in the basin-periphery zone of least deflection. Numerical details of the model are tabulated in and discussed by Currey and Burr (1988).

hypsonetric and chronometric calibration, and refinements in its numerical structure to better reflect hydrologic and tectonophysical nonlinearities, the model will evolve into an even more versatile and robust tool in the future.

Great Salt Lake Hypotheses

Several important hypotheses regarding spatiotemporal patterns of Great Salt Lake after about 13,000 yr B.P., i.e., during marine isotope stage 1 (Shackleton and Opdyke, 1973) and marine chronostratigraphic interval 2.0-1.0 (Martinson et al., 1987), have resulted from recent research. The Bonneville paleolake cycle terminated abruptly between 13,000 and 12,000 yr B.P. (Figure 12), which was coincident with termination I of the deep-sea record (Broecker and van Donk, 1970). The final stages of the regression to the pre-Gilbert low (Figure 12, PGL) are marked by red beds at many localities near Great Salt Lake (Figure 16) and around the Great Salt lake Desert (Currey et al., 1988a). The red beds probably were derived from FeS_2 -bearing anoxic facies of Lake Bonneville deep-water (maximum depth ≥ 370 m) sediments and probably were reddened by oxidation as receding brines reworked them basinward across mudflats at the margins of the dwindling lake. The offshore correlative of the red beds is the base of a thick sequence of interbedded mirabilite ($\text{Na}_2\text{SO}_4 \cdot 10\text{H}_2\text{O}$) and mud which underlies the deepest part of Great Salt Lake (Eardley, 1962a; Mikulich and Smith, 1974), and which is currently an object

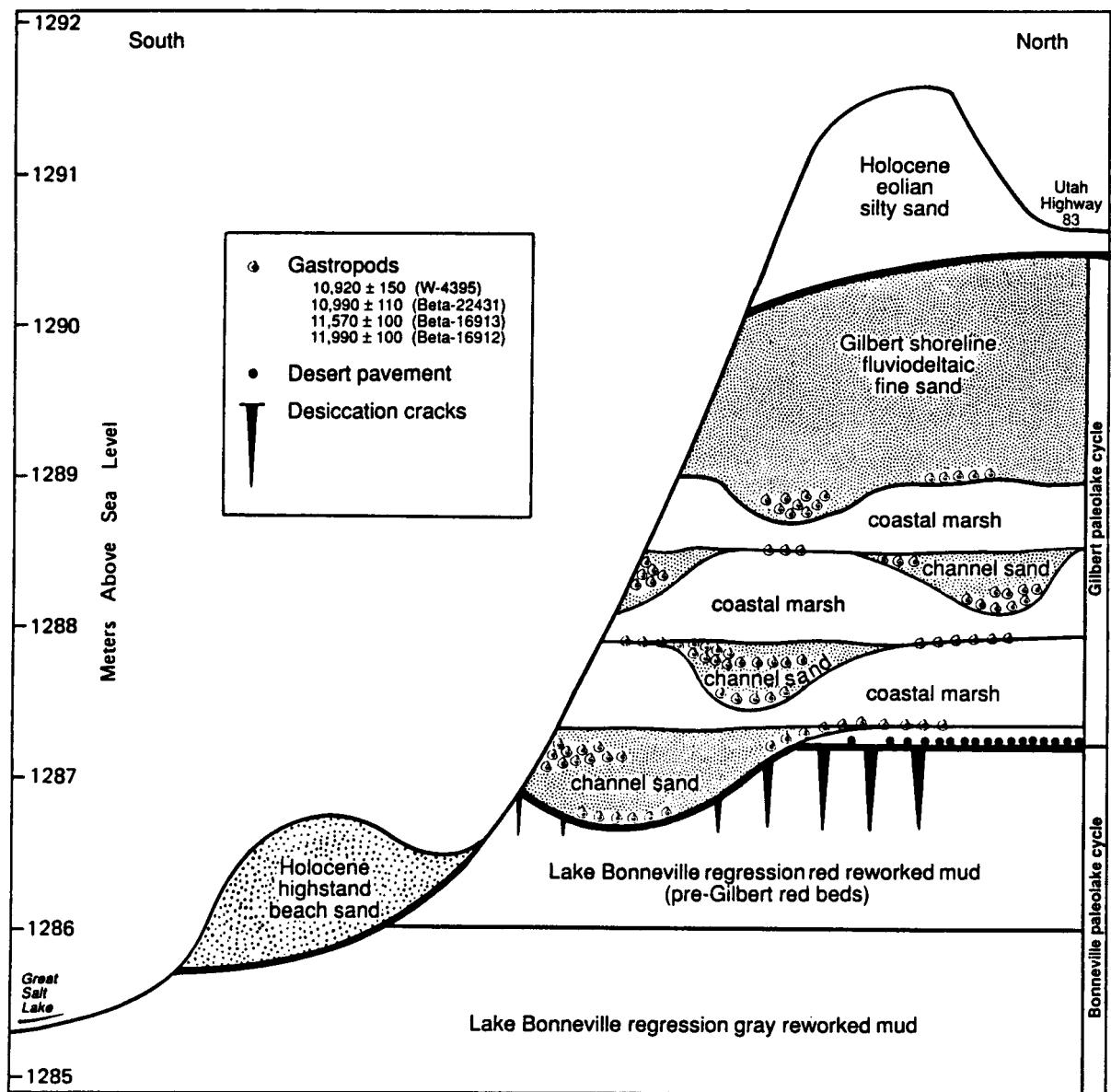


Figure 16. Schematic cross section of late Quaternary stratigraphy near the northeast shore of Great Salt Lake (adapted from Smith et al., 1989, fig. 64), 10 to 20 km west of Corinne (Figure 21).

of increasing spatial and temporal resolution (Currey, 1988b, table 1).

In the Great Salt Lake and Great Salt Lake Desert subbasins of the Bonneville basin, several radiocarbon ages suggest that the highest stage of Great Salt Lake culminated at the Gilbert shoreline complex (Figure 12, GSC) between 10,900 and 10,300 yr B.P., in latest Pleistocene time. In the Carson Desert of the Lake Lahontan system of subbasins, a minor paleolake cycle which was similar in magnitude to the Gilbert cycle culminated at what has been termed the Russell shoreline, at a time which two radiocarbon ages suggest was about 11,100 yr B.P. (Currey, 1988a); a third radiocarbon age, on *Anodonta* at the classic Humboldt Bar (Russell, 1885, pl. XVIII) sill between the Carson Desert and Humboldt Sink subbasins of the Lahontan basin, suggests that the Russell shoreline development there continued until at least 10,400 yr B.P. (Beta-29024). Indices of comparative chronometry are sometimes helpful in comparing ages which have been obtained from paleolake materials. An index which can be used to compare two ages before present—of two samples from one paleolake basin or of one sample from each of two paleolake basins—is the *comparative heterochronology index (chi)*, where $\chi = |age_1 - age_2| \div 0.5(age_1 + age_2)$. As measured by this index, within-basin and between-basin paired radiocarbon ages of suitable materials from familiar Gilbert and Russell stratigraphic contexts in the Bonneville and Lahontan basins are typically homochronous or nearly so (Table V).

TABLE V

Comparative heterochronology index (χ)

Descriptor	Chi (χ) range
Homochronous	$0 \leq \chi < 0.02$
Homeochronous	$0.02 < \chi < 0.2$
Heterochronous	$0.2 < \chi \leq 2$

Other indices treat comparative paleolake morphometry, and can aid in making direct size comparisons of paleolakes such as the Gilbert and Russell water bodies. Two such indices are the *paleolake height index (phi)*, where $\phi = (\text{water depth}) \div (\text{greatest water depth during isotope stage 2})$, and the *paleolake surface index (psi)*, where $\psi = (\text{water area}) \div (\text{greatest water area during isotope stage 2})$. For the Gilbert water body $\phi = 0.160$ (Figure 17) and $\psi = 0.333$ (Figure 18), and for the Russell water body $\phi = 0.163$ and $\psi = 0.331$. This comparison suggests that the terminal Pleistocene paleolakes in the Bonneville basin and Carson Desert, which are 500-600 km apart (Figure 1), were as similar morphometrically as they were chronometrically (Currey, 1988a).

Elsewhere in the Great Basin, a similar paleohydrographic pattern may be present in many of the smaller subbasins, where the last perennial lakes appear to have occurred in terminal Pleistocene, rather than Holocene, time. Even in the northwestern extremity of the Great Basin, ponds, marshes, and perennial streams in central Oregon seem to have been more numerous between 11,000 and 10,000 yr B.P. than at any time subsequently (Bedwell, 1973, fig. 10). The probable timing of the very similar Gilbert and Russell transgressions, and the possible timing of the last lakes in smaller subbasins, suggests that the northern hemisphere summer insolation maximum which resulted from the earth's orbital elements about 11,000 yr B.P. (Berger, 1978) could have been an underlying factor. It is tempting to hypothesize that general circulation driven by the summer insolation maximum delivered unusually large

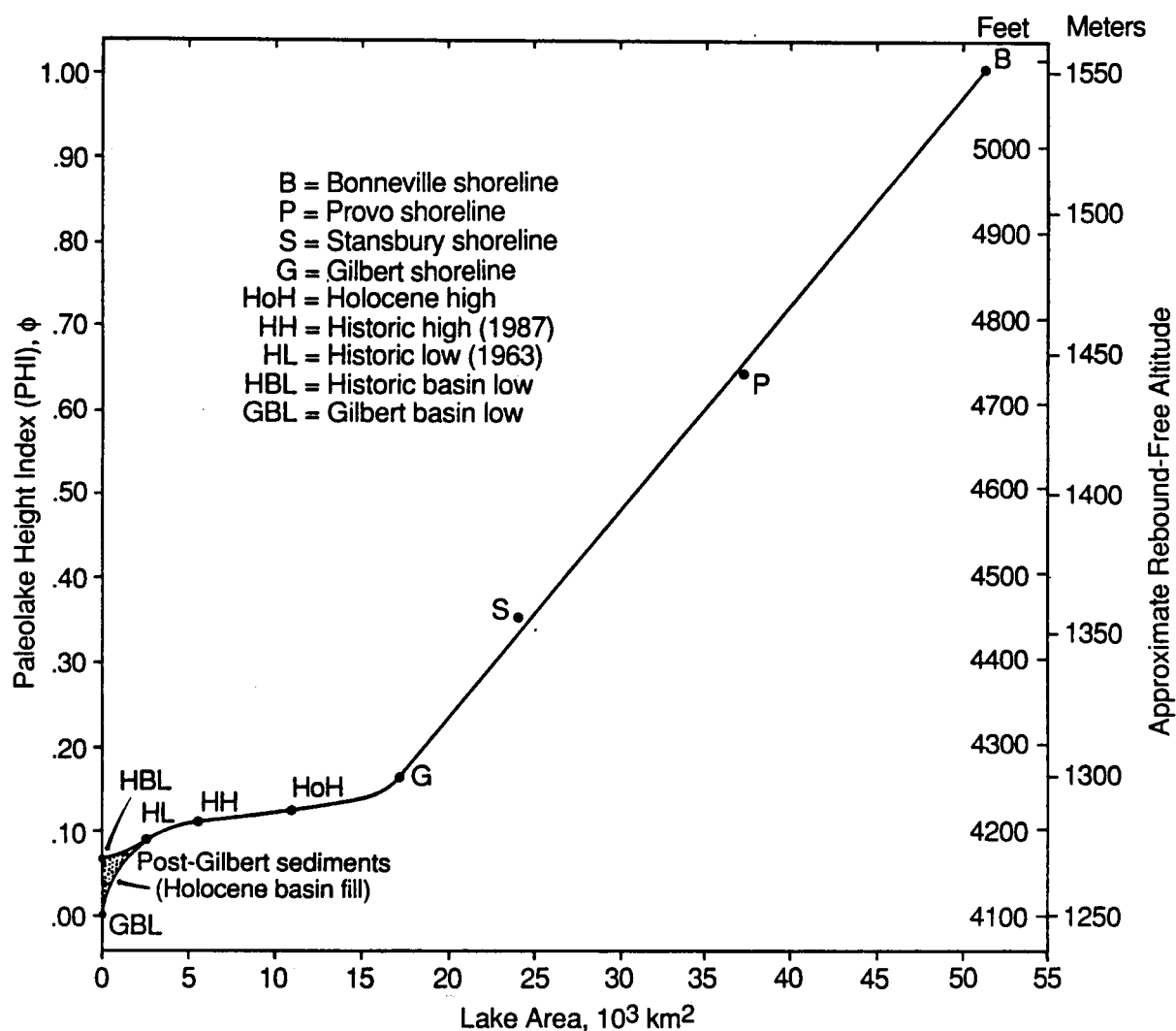


Figure 17. Generalized hypsographic curve of the Bonneville closed basin, showing relations among selected late Quaternary lake and basin-floor levels (adapted from Currey and Oviatt, 1985, fig. 4). Rebound-free altitudes are approximated basin wide using a normalizing equation employed by Currey and Oviatt (1985, p. 1087). The dimensionless phi index (ϕ) is scaled from zero at the lowest basin floor of terminal Pleistocene (Gilbert shoreline) age to 1.00 at the highest late Pleistocene (Bonneville) shoreline.

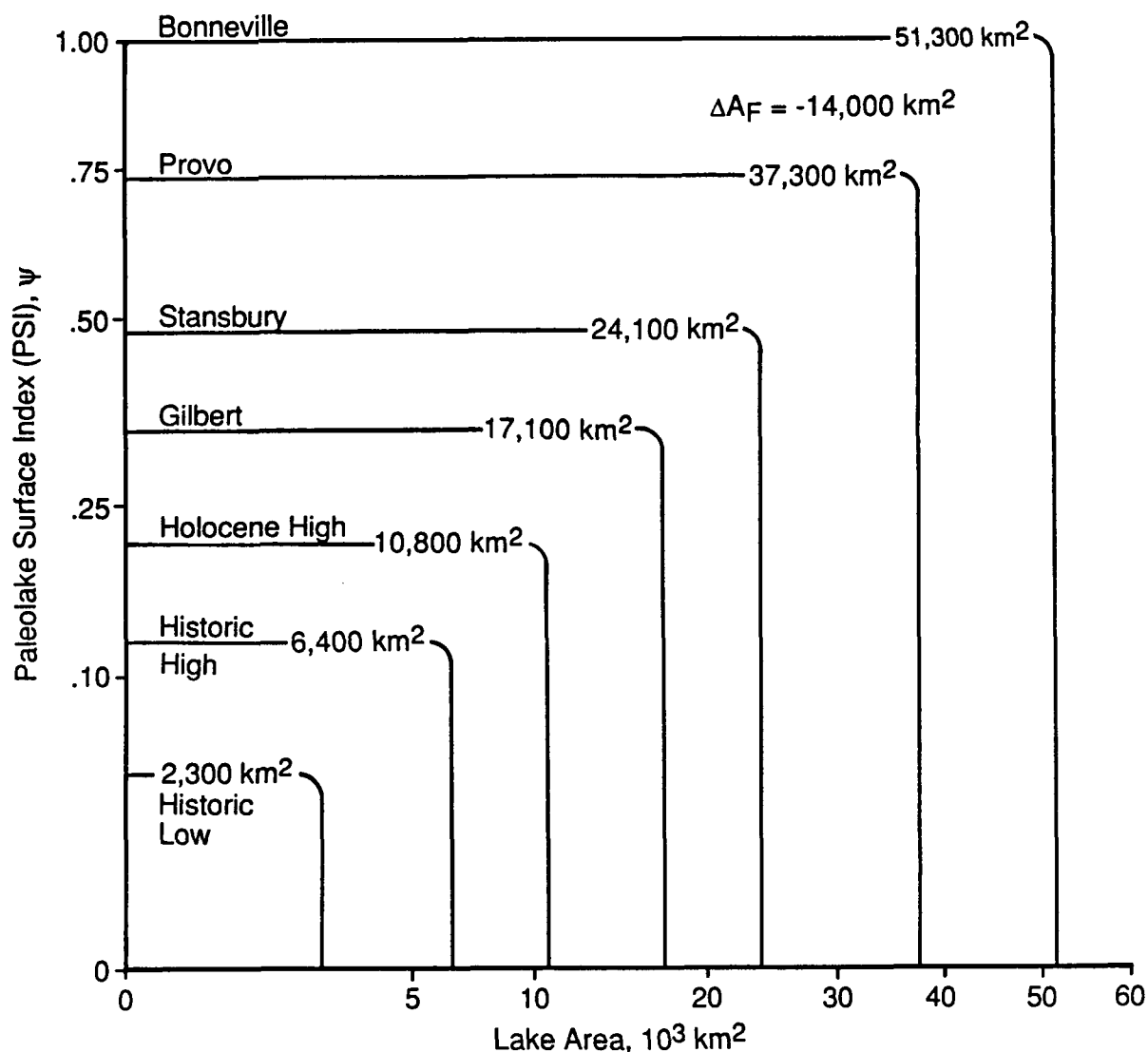


Figure 18. Proportional-area representation of selected late Quaternary water surfaces in the Bonneville basin. Surface area was reduced 14,000 km² by the catastrophic Bonneville Flood. The dimensionless psi index (ψ) is scaled from zero in the case of complete basin-floor reliction—which was seldom, if ever, possible in the Bonneville basin and other semidesert basins with significant areas of runoff-producing highlands—to 1.00 at the most expansive late Pleistocene (Bonneville shoreline) stage.

quantities of tropical moisture to the northern Great Basin during Gilbert-Russell time. That hypothesis is not inconsistent with climatic modelling and packrat (*Neotoma*) midden data which suggest that monsoonal flow and summer convective precipitation increased substantially in the southern Great Basin between 12,000 and 8,000 yr B.P. (Spaulding and Graumlich, 1986).

Historically, the Great Basin has been transitional between a region of predominantly winter precipitation to the west and a region of predominantly summer precipitation to the east (Figure 19). Prehistorically, during the terminal Pleistocene insolation maximum, a 6° westward shift of the western U.S. moisture seasonality boundary would have brought increased summer cloud cover, humidity, and rainfall to all of the Great Basin, including the semidesert basins of Oregon. Perennial water bodies throughout the northern Great Basin probably would have been sustained more effectively under those conditions than under the present regime. However, a 6° westward shift would have placed the moisture seasonality boundary near what is now the eastern limb of a subtropical high, which is a region of atmospheric subsidence and divergence, and hence dryness, during the summer. Therefore, if a significant westward shift of the moisture seasonality boundary occurred during the insolation maximum, a weaker and/or displaced subtropical high is implied.

Holocene stages of Great Salt Lake fluctuated within a relatively narrow range (Figure 12, HS). The upper envelope of static water during Holocene time was 1,286.7 m above sea level

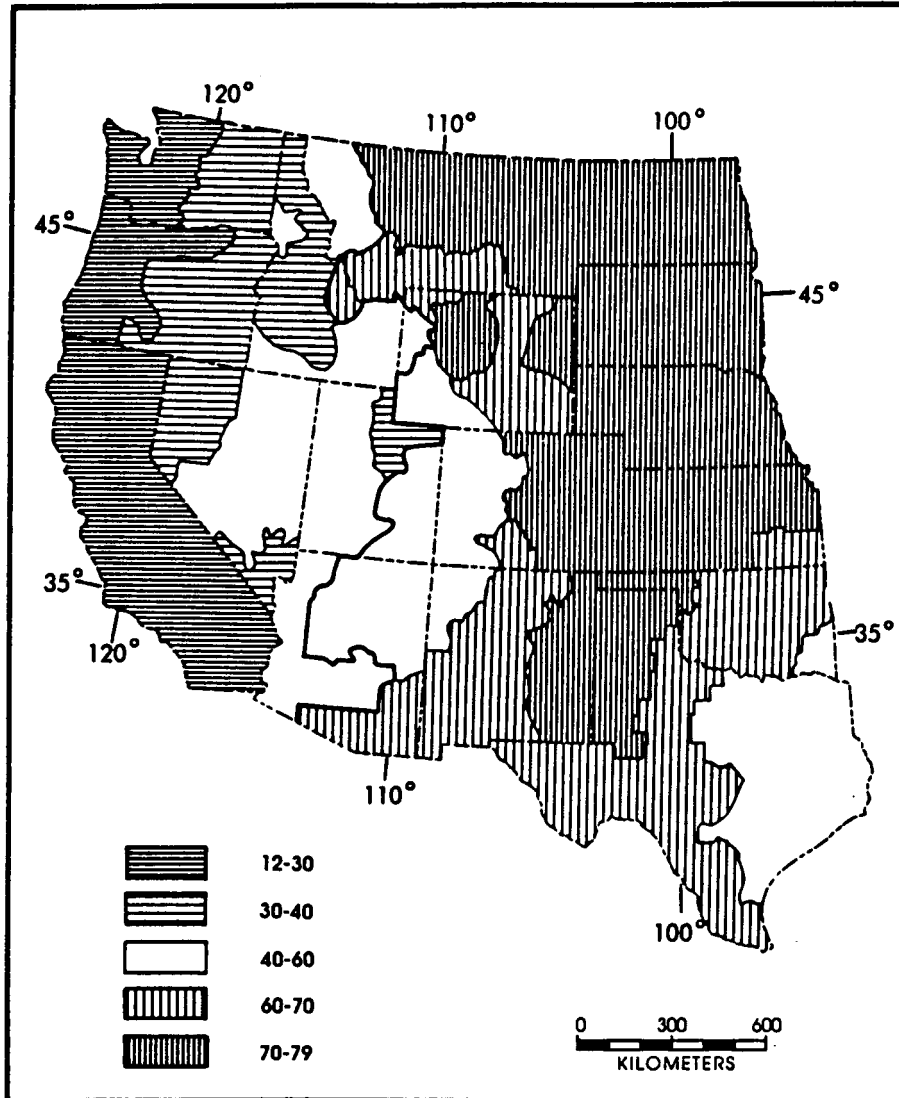


Figure 19. Summer (April through September) precipitation in the contiguous western United States as a percentage of annual precipitation, based on 1931-1960 monthly averages for 136 U.S. Weather Bureau climatic divisions (after Currey, 1976, fig. 3). Historically, on a subcontinent scale, the 50-percent isoline (western U.S. moisture seasonality boundary) coincides very closely with the 112° W. meridian.

(Currey et al., 1988b), which was only about 3 m above the historic high stage and only about 6 m above the historic average stage (Arnow, 1984). The highest Holocene bluffs which were cut in soft sediments by exceptionally dynamic water—by storm surges and perhaps by seismic seiches—are locally as much as 6 m higher than the highest Holocene microberms which were built at low-energy localities (Merola et al., 1989). Radiocarbon ages of shoreline geomorphic features constrain the highest Holocene stage to between about 7,100 and 1,400 yr B.P. (Currey et al., 1988a). Ages of lakeshore archaeological sites also constrain the highest Holocene stage to before about 1,400 yr B.P. (Currey and James, 1982, pp. 40-42). Age-calibrated environmental reconstructions using stable isotopes and carbonate geochemistry (Grey and Bennett, 1972, pp. 11-14; McKenzie and Eberli, 1987, fig. 2) and palynology (Mehring, 1985, figs. 10 and 11) of lake-bottom cores suggest that the highest Holocene stage of Great Salt Lake occurred between 3,000 and 2,000 yr B.P. The highest Holocene stage in the Sevier Lake subbasin of the Bonneville basin, 180 km south of Great Salt Lake, probably occurred about 2,600 yr B.P. (Oviatt, 1988, p. 17).

The probable timing of the highest Holocene stage suggests that it may have been the lacustrine equivalent of a widely reported possible episode of Neoglaciation (Porter and Denton, 1967, fig. 4; Denton and Karlén, 1973, fig. 1; Grove, 1988, fig. 10.23). The terms *Neopluvial* and *Neolacustral* have been used to describe late Holocene phases of increased moisture availability in the northwestern (Allison, 1982, pp. 70-71) and northeastern

(Currey and James, 1982, p. 29) Great Basin. Landmark late Holocene stages of Great Salt Lake, which has inundated the lowest part of the northeastern Great Basin continuously since the Bonneville paleolake cycle, are depicted in Figure 20. The areal extent of the highest Holocene static water (1,286.7 m) in the Great Salt Lake region is delineated in Figure 21.

Concluding Observations

A series of recurring themes have been observed in the paleolake record of the Great Basin, and are helping to guide paleolake research in that region.

(1) One of the most vexing problems in paleolake studies has been that of bridging the correlation gap between littoral and pelagial records (e.g., Eardley and Gvosdetsky, 1960, p. 1343), particularly in sedimentary basins which have complex lateral and vertical facies changes (e.g., Miall, 1981). To solve stratigraphic problems which involve basin-wide facies architecture, each site-specific study must be regarded as more than an independent module of stratigraphic analysis. Site-specific studies must also be regarded as interdependent modules of stratigraphic synthesis, i.e., as nodes in networks which link basin architecture and facies architecture on basin-wide (or subbasin-wide), cycle-deep spatial and temporal scales (e.g., Russell, 1885; Gilbert, 1890; Eardley et al., 1957; Eardley, 1962b; Benson, 1978; Spencer et al., 1984; Oviatt, 1988; Sack, 1989b).

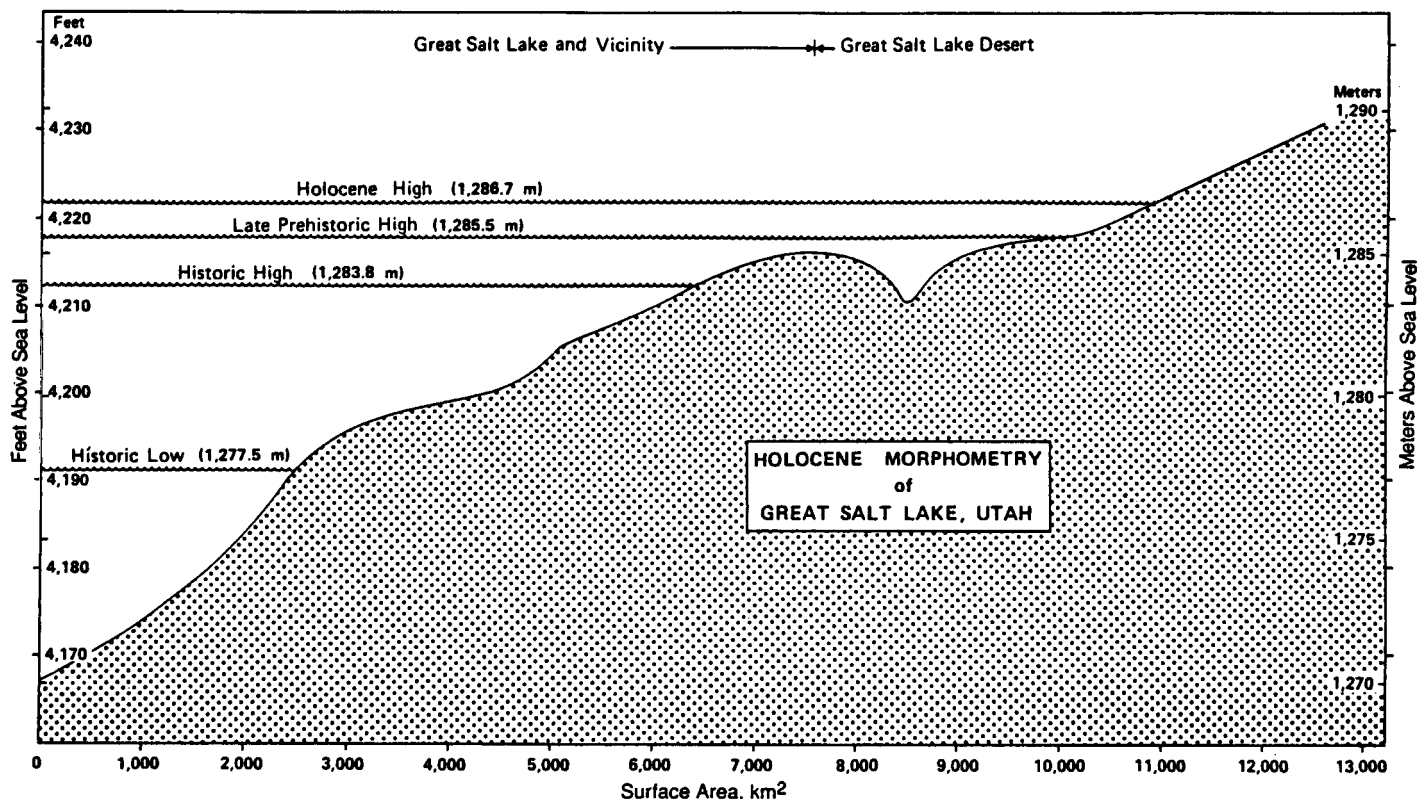


Figure 20. Hypsographic curve of the lowest part of the Bonneville basin, showing singular levels of Great Salt Lake (Currey, 1987, fig. 4). The historic low occurred in 1963, the historic high in 1873 and 1986-87, the late prehistoric high probably about A.D. 1700, and the Holocene high probably between 3,000 and 2,000 yr B.P. Prior to 1987, local runoff ponded naturally in the Great Salt Lake Desert depression during wet intervals. In 1987, a State of Utah pumping project began to circulate Great Salt Lake surface brine through the Desert depression, to increase evaporative losses from the hydrologic basin after an exceptionally wet interval had resulted in unusually high lake levels.

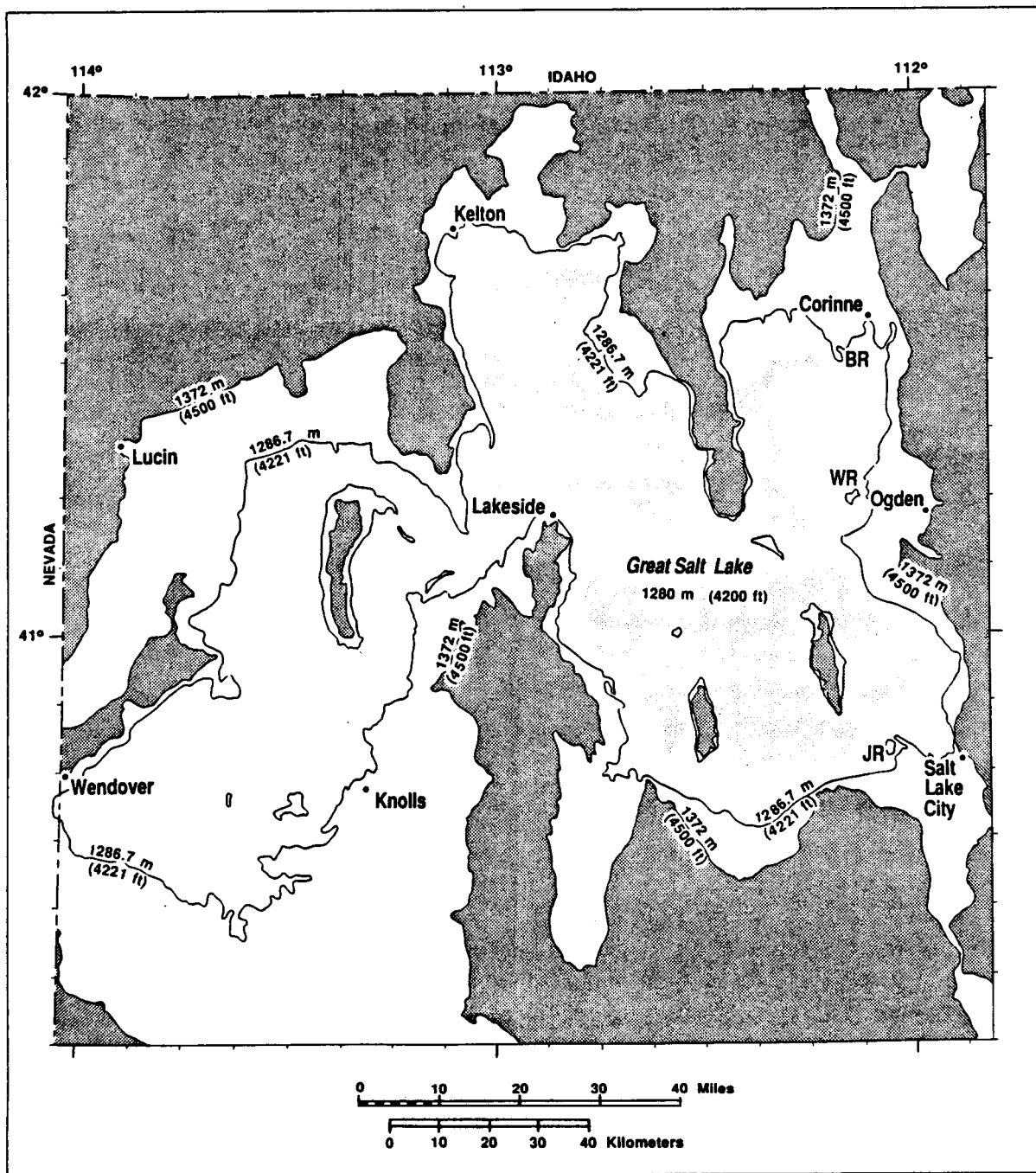


Figure 21. Map of the Great Salt Lake region (Merola et al., 1989, fig. 9) depicting the maximum extent (10,900 km²) of standing water during the Holocene highstand, when the upper limit of static water was 1,286.7 m above sea level. Holocene highstand paleodeltas of the major inflowing streams are: BR = Bear-Malad river system, WR = Weber and Ogden rivers, and JR = Jordan-Provo river system.

In other words, paleolake-wide stratigraphic synthesis, with its synergistic qualities, is the larger context within which stratigraphy is analyzed at any site. Conversely, site-specific stratigraphic analysis is the indispensable means by which the predictive power of paleolake-wide synthesis is tested and refined.

(2) The full geologic expression of a paleolake cycle comprises the full lateral and vertical range of the cycle's depositional sequence (Mitchum et al., 1977), or *alloformation* (North American Commission on Stratigraphic Nomenclature, 1983, pp. 865-867). A depositional sequence (alloformation) is bounded at its base and top by unconformities or, at continuously inundated localities, by conformities which are correlative with unconformities in littoral areas.

(3) Quantitative syntheses of basin-wide facies architecture and neotectonic architecture require paleolake data which are georeferenced to x, y, and z spatial coordinates, preferably in the UTM planimetric system and with sub-meter resolution of the vertical coordinate. Geodetic total station (GTS), global positioning system (GPS), and geographic/land information system (GIS/LIS) technologies are well suited to these applications.

(4) *Limneotectonics* (lacustrine neotectonics)—the use of Quaternary paleolake datums (Currey, 1988b), each with a unique combination of paleolimnologic, morphometric, and chronometric properties, as long-baseline *tiltmeters* (e.g., Rose, 1981) and site-specific *slipmeters*—can be effective in reconstructing the neotectonic evolution of many semidesert basins.

(5) The depositional littoral record has long been regarded as an important source of lacustrine information (Gilbert, 1885). Unfortunately, at many classic and potentially classic localities, littoral (and other geologic) antiquities which are irreplaceable as scientific and educational resources are being rapidly consumed by a growing demand for mineral and land resources.

(6) Littoral deposits are important as gauges, or "dipstick" marks, which often record hypsometric (Figure 22) and sometimes chronometric (e.g., Easterbrook, 1988) dimensions of paleolake hydrographs (e.g., Currey and Oviatt, 1985, fig. 3). Furthermore, littoral deposits provide data which are important in the reconstruction of nearshore and coastal environments, including wave energy patterns and sediment budgets (e.g., Carter, 1988, pp. 192-244). In selected cases, littoral deposits are important as paleolake datums (Figure 9, *PLL*) in limneotectonic analysis (Currey, 1988b). At some basin-flank localities, littoral deposits have endured to provide shoreline evidence of paleolake cycles which predate the most recent deep-lake cycle (e.g., McCoy, 1987; Oviatt and Currey, 1987; Scott, 1988).

(7) The littoral record of a major paleolake cycle tends to be very asymmetric in its information content, with far more information dating from the transgressive phase than the regressive phase. In part, that is because transgressive phases have tended to be more protracted. More importantly, however, it is because transgressive phases have tended to be very effective in transforming subaerial materials into littoral deposits, i.e., in

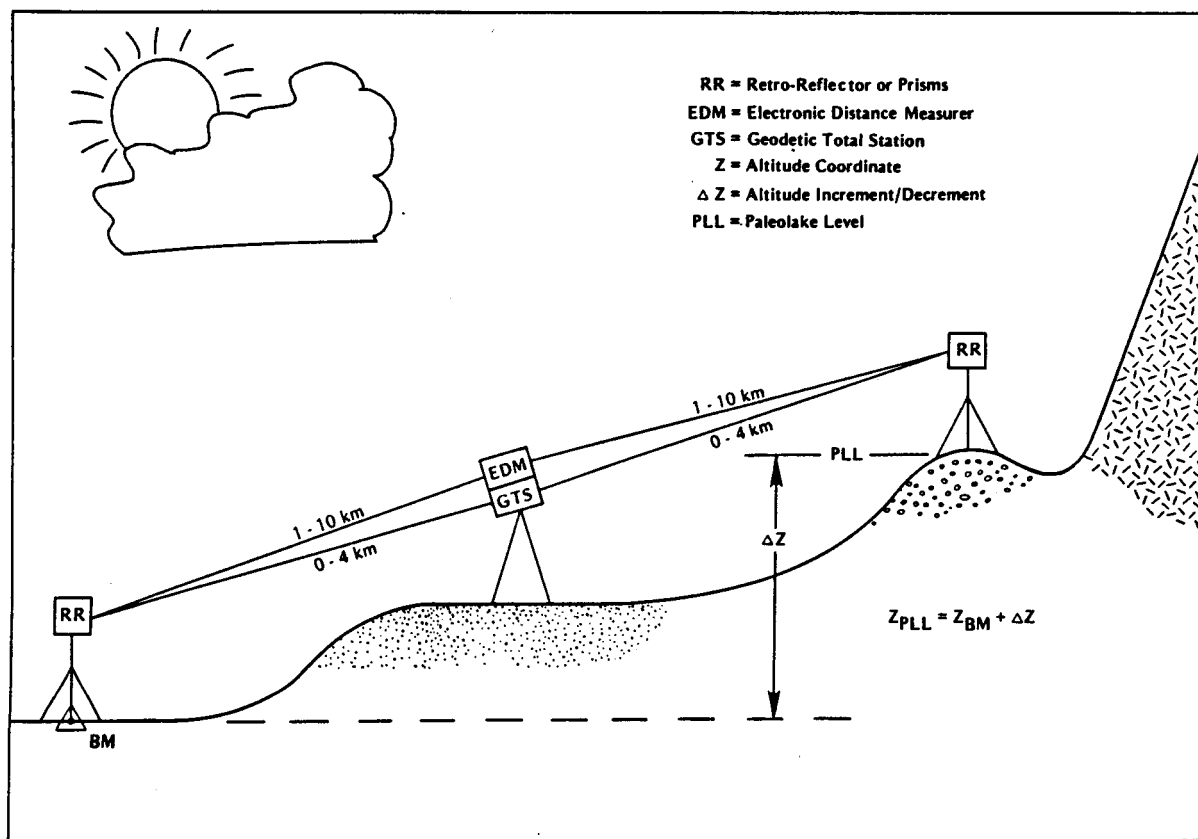


Figure 22. Horizontal and vertical coordinates of paleolake field sites are readily obtained from networks of known or assumed control points using the theodolite, infra-red distance meter, and micro-processor capabilities of a geodetic total station (GTS).

bringing pre-lacustrine surficial geology into equilibrium with lacustrine processes.

(8) Notable exceptions to (7) occur in many deltaic depocenters, where information stored in descending flights of offstepping subdeltas during regressive phases rivals that stored during transgressive phases.

(9) Where littoral sediments provide high-resolution morphostratigraphic records of lacustrine variation, evidence of long-term lake-level change during transgressive or regressive phases commonly is overprinted with evidence of shorter-period (higher-frequency) lake-level oscillations. Higher-frequency signals are particularly evident in the internal and external structure of beaches, and in the structure and mineralogy of littoral carbonates. Efforts to assess the regional coherence and origins of higher-frequency signals are of growing importance in Great Basin paleolake studies.

(10) Bayhead barriers, baymouth barriers, cusate barriers, spits, tombolos, and other well-defined bodies of littoral sediment (Zenkovich, 1967, pp. 383-447) commonly represent on the order of one *gigasecond* ($1 \text{ Gs} = 31.7 \text{ y}$) of variable-rate onshore and longshore geomorphic work (Currey and Burr, 1988). Such bodies commonly are clustered spatiotemporally in shoreline complexes which represent longer intervals of sustained variable-rate geomorphic work.

(11) Major (Gs-scale) stages of closed-basin lakes comprise many minor (intra-annual and inter-annual) fluctuating stages.

Repeated water-level deviations above and below median levels of major stages create opportunities for depositional shorelines to undergo cumulative aggradation, and for erosional shorelines to undergo cumulative degradation. Consequently, an erosional shoreline tends to reflect the lower geomorphic envelope of a major stage. A *drift-aligned* depositional shoreline at a locality with unimpeded longshore transport (Davies, 1980, p. 136) tends to reflect the upper static-water envelope of a major stage. A *swash-aligned* depositional shoreline at a locality with impeded longshore transport (Davies, 1980, p. 135) tends to reflect the upper limit of dynamic water, which can be several meters higher than contemporaneous static water. Super-elevation of swash-aligned shorelines tends to be greatest where pocket beaches maximize onshore surging of breaking waves.

(12) From the standpoint of paleolake information, the erosional littoral record has long been regarded as a vacuity which is of greater scenic than scientific value. However, at localities with suitable lithologies and erosional histories, recent advances in rock-surface geochronology and paleoecology (e.g., Dorn, in press) are starting to clarify hitherto intractable dimensions of the erosional record.

(13) Outside of basin-floor areas, including deltas and estuaries, only minute fractions of the offshore and nearshore areas which were inundated during the last deep-lake cycle are likely to retain a portion of their pelagial sediment cover. Most basin-flank pelagial sediments were reworked long ago, by being

washed onto basin floors or blown onto downwind terrain.

(14) Rates of sedimentation in paleolakes clearly varied significantly in space and time. Some of that variation was inherent in littoral and fluviolacustrine sedimentation, some resulted from sediment reworking during and following regressive stages, and some has resulted from geomorphic rejuvenation of sedimentation by basin-floor and basin-flank neotectonics (Table II).

(15) Isochronous or quasi-isochronous marker beds can be of crucial importance in correlating within and among paleolake basins. At least a dozen tephra units (Oviatt and Nash, 1989, fig. 10) and distinctive lithofacies markers of late Quaternary age are currently in local or regional use in the Bonneville basin. Tephra layers have been used even more widely as marker beds in the western Great Basin (e.g., Davis, 1985).

(16) Paleolake records in systems of formerly interconnected subbasins—such as the Lake Lahontan system of subbasins (Benson and Thompson, 1987b, fig. 1 and table 1)—tend to contain less coherent information, but more total information, than do paleolake records in less complex hydrographic systems.

(17) Resolving the spatiotemporal and paleoenvironmental dimensions of low-amplitude Holocene fluctuations is not inherently simpler, or more trivial, than resolving similar dimensions of higher-amplitude Pleistocene fluctuations (e.g., Berglund, 1986; Logan, 1987; Last and Slezak, 1988).

(18) As Mifflin and Wheat (1979, pp. 15 and 27) and Quade and Pratt (1989) have demonstrated so clearly, care must always be taken to differentiate between lacustrine and palustrine records in what sometimes are paleommarsh or paleoplaya basins, and not paleolake basins at all. Moreover, what appear to be basins with hydrographic closure are sometimes deceiving—features such as wet meadows and beaded strings of playas are just as likely to occur on the floors of hydrographically open semi-bolsons (Peterson, 1981, pp. 30-34). Paleohydrologic, paleoclimatic, and tectonophysical reconstructions which are based on the geographical distribution and morphometry of paleolakes are vulnerable to potential errors in paleolake recognition and delineation.

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